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TRANSACTIONS OF THE CENTRAL AEROLOGICAL OBSERVATORY
(SELECTED ARTICLES)

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THE INVESTIGATION OF CERTAIN ERRORS IN THE DETERMINATION OF THE ABSORPTION OF SOLAR RADIATION IN THE ATMOSPHERE WITH PYRANOMETERS

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As is apparent from Reference [1], the observations of the absorption of solar radiation in the lower troposphere made in 1949-55 at a number of locations in the Soviet Union led to conclusions, some of which are rather difficult to explain. Consequently we are forced to make the assumption that the different types of errors in the determination of absorption by the given method are in fact substantially greater than had heretofore been assumed. This paper supplements the investigation of errors contained in References [2] and [3] and gives a more detailed analysis of those errors which were not taken into consideration earlier or were not analyzed in sufficient detail.

1. The Effect of the Dependence of Sensitivity of Pyranometers on the Incidence Angle of Radiation

A. Theory

The effect of the dependence of the sensitivity of pyranometers on the incidence angle of radiation has already been considered in Reference [2]. The calculation given in this work for two Yanishevskiy pyranometers gave a negligibly small magnitude. However, it was assumed in this calculation that only the ratio between direct, scattered, and reflected radiation changes with altitude over the earth's surface, while the distribution of brightness over the sky and lo... hemisphere does not noticeably change. In the calculations

to be given below, this assumption was discarded. In addition, the calculations were made for a greater number of instruments and under various conditions of observation.

Since there is little aerosol-particle absorption in all regions of the solar spectrum, in calculating the indicated errors it was assumed that pure absorption takes place only in water-vapor bands.

The subscript 1 will denote quantities referring to the lower pyranometer, and the subscript 2 will denote quantities referring to the upper pyranometer. Further, in accordance with the methods of treating observations of absorption of solar radiation used in GMS observatories, we will assume that they directly use the conversion factors of both pyranometers obtained in calibrating by the sun-shadow method at a certain incidence angle α_0 of solar beams, without introducing any corrections for the dependence of the sensitivity on the incidence angle. In the majority of cases the pyranometers were calibrated at normal beam incidence ($\alpha_0 = 0$). Let $\psi_1(\alpha), \psi_2(\alpha)$ be the ratios of the sensitivity of the lower and upper pyranometers at incidence angle α to their sensitivity at incidence angle α_0. Then the density of the ascending and descending radiation flows, obtained in processing, can be expressed by the following formulas:

$$E_1' = \int_{\lambda=0}^{\infty} \int_{4\pi} i_\lambda \cos\theta \, \psi_1(\theta) \, d\omega \, d\lambda, \tag{1}$$

$$E_2' = -\int_{\lambda=0}^{\infty} \left[s_\lambda \cos\theta_\odot \, \psi_2(\pi - \theta_\odot) + \int_{4\pi} i_\lambda \cos\theta \, \psi_2(\pi - \theta) \, d\omega \right] d\lambda. \tag{1'}$$

Here Θ denotes the angle between the direction of the rays and the axis Z directed upward, so that for the rays proceeding from the lower hemisphere, $\cos\Theta > 0$, while for rays proceeding from the upper hemisphere, $\cos\Theta < 0$. Let s_λ and i_λ denote the spectral densities of direct solar radiation and the intensities of the diffuse light pro-

ceeding from the lower and upper hemispheres. The integration over the solid angle in the right side of Eqs. (1) and (1') extends not over the hemisphere to which the pyranometer is directed, but rather over the entire sphere, since all the pyranometers possess a certain sensitivity for incidence angles >90°* because of the reflection of radiation from the inner surface of the glass bell. Practically speaking, it is necessary to integrate up to $\alpha = 115°$. For higher values of α, we can assume that $\psi_1(\alpha) = \psi_2(\alpha) = 0$.

From Eqs. (1) and (1'), we obtain the following expression as the result of the calculation of the absorption intensity:

$$q' = \frac{d(E_2' - E_1')}{dz} = -\int_{\lambda=0}^{\infty}\left\{\frac{ds_\lambda}{dz}\cos\theta_\odot\psi_1(\pi-\theta_\odot) + \right.$$

$$\left. + \int_{4\pi}\frac{di_\lambda}{dz}\cos\theta\,[\psi_2(\pi-\theta)+\psi_1(\theta)]\,d\omega\right\}d\lambda, \qquad (2)$$

The true absorption-intensity magnitude, evidently, is the following:

$$q = -\int_{\lambda=0}^{\infty}\left\{\frac{ds_\lambda}{dz}\cos\theta_\odot + \int_{4\pi}\frac{di_\lambda}{dz}\cos\theta\,d\omega\right\}d\lambda. \qquad (2')$$

From (2) and (2') it is not difficult to obtain the following expression for the unknown error,

$$\delta' q = q' - q = q\,[\psi_1(\pi-\theta_\odot) - 1] + \int_{\lambda=0}^{\infty}\int_{4\pi}\frac{di_\lambda}{dz}\cos\theta\, f(\theta)\,d\omega\,d\lambda, \qquad (3)$$

where

$$f(\theta) = \psi_1(\pi-\theta_\odot) - \psi_2(\pi-\theta) - \psi_1(\theta). \qquad (3')$$

The first term on the right side of Eq. (3),

$$\delta_1' q = q\,[\psi_1(\pi-\theta_\odot) - 1] \qquad (4)$$

produces an error in the determination of q because of inaccuracy in the conversion factor of the upper pyranometer for direct radiation. It is easy to see that this error never becomes substantial. Actually, the absorption measurements are always taken at high solar altitudes, for which the relative sensitivity of the Yanishevskiy

pyranometers deviates from 1 only by a small per cent. Consequently, $\delta_1' q$ should be many times less than the accidental errors which undoubtedly are tens of per cent of $\underline{q}$. In addition, $\delta_1' q$ is completely eliminated if the upper pyranometer is calibrated in a horizontal position approximately at the same solar altitude at which the observations were made, or if, in processing, we introduce into the pyranometer readings corrections corresponding to the solar altitude.

To make it easier to determine the second term on the right side of Eq. (3), we will assume that the solar spectrum can be divided into two regions (not necessarily contiguous), that is to say, a short-wave region (I) in which the radiation is attenuated only because of the scattering and a long-wave region (II) of water-vapor absorption bands whose scattering can be neglected. It is easy to see that the second term in Eq. (3) for the long-wave region

$$\delta_2' q = \int\limits_{II} d\lambda \int\limits_{4\pi} \frac{dI_\lambda}{dz} \cos\Theta f(\Theta)\, d\omega \tag{5}$$

is negligibly small. Actually, in accordance with our assumptions, the brightness of the sky in this region and the air haze are equal to zero at all altitudes and, consequently, when we integrate over ω in Eq. (5), we can limit ourselves to the lower hemisphere. However,

$\int\limits_{II} d\lambda \int \frac{dI_\lambda}{dz} \cos\Theta\, d\omega$, integrating with respect to ω over the lower hemisphere, is obviously the absorption intensity of the radiation reflected from the earth in the water-vapor bands – a very small quantity approximately one order less than the intensity of the absorption by water vapor of direct solar radiation. Since the coefficient in Eq. (5) $f(\Theta)$ is also small for almost all directions, there can be no doubt that we can completely neglect $\delta_2' q$. We also checked this conclusion by very detailed calculation.

For the short-wave region of the spectrum, the second term on

the right side of Eq. (3) is transformed by replacing di_λ/dz with $di_\lambda/d\tau_\lambda \; d\tau_\lambda/dz$, where τ_λ is the optical density of that region of the atmosphere which lies beneath the level being considered and by introducing into the numerator and denominator the normal illumination of direct solar radiation s_λ. Here we obtain the following:

$$\delta_3' q = \int s_\lambda \frac{d\tau_\lambda}{dz} \int_{4\pi} \frac{1}{s_\lambda} \frac{di_\lambda}{d\tau_\lambda} \cos\Theta f(\Theta)\, d\omega\, d\lambda. \tag{6}$$

If we limit ourselves to the first approximation at which we do not consider multiple scattering and attenuation of primarily scattered radiation, $1/s_\lambda \; di_\lambda/d\tau_\lambda$ will be a function of λ only because the relative scattering indicatrix is a function of λ. Since $1/s_\lambda \; di_\lambda/d\tau_\lambda$ in Eq. (6) is under the integral sign over the solid angle, as a result of this, the dependence on λ is leveled to an even greater extent. Consequently, it is permissible to substitute $1/s_\lambda \; di_\lambda/d\tau_\lambda$ in (6) by its average value over λ: $1/s \; di/d\tau$. It is further apparent that

$$s_\lambda \frac{d\tau_\lambda}{z} = S_\lambda^* e^{-(\tau^* - \tau) m_\odot} \frac{d\tau_\lambda}{dz} = \frac{1}{m_\odot} \frac{ds_\lambda}{dz},$$

where $m_\odot$ is the optical mass in the sun's direction, while the asterisk will here and everywhere denote that the magnitude under consideration refers to the upper boundary of the atmosphere. Consequently, the expression for $\delta_3' q$ can be given the following form:

$$\delta_3' q = \frac{1}{m_\odot} \frac{dS_1}{dz} \tilde{f}, \tag{7}$$

where

$$S_1 = \int s_\lambda \, d\lambda, \tag{8}$$

$$\tilde{f} = \int_{4\pi} \frac{1}{s} \frac{di}{d\tau} \cos\Theta f(\Theta)\, d\omega. \tag{9}$$

The coefficient $\tilde{f}$ can be considered as some weighted-average value of the function $f(\Theta)$. Actually, it follows in the first approximation from the so-called radiation-transfer equation

$$\frac{1}{s}\frac{di}{d\tau}\cos\Theta = \frac{\gamma}{4\pi},$$

where γ is the coefficient of the directed light scattering, normalized according to the law $\int_{4\pi}\frac{\gamma}{4\pi}\,d\omega = 1$. Consequently, the expression 1/s di/dτ cos Θ can be considered as a weight with which we average the function f(Θ). It is interesting to note a certain analogy between Eqs. (4) and (7). The right sides of both equations are the products of the coefficient determining the deviation of the pyranometer sensitivity from the isotropism by the attenuation of solar radiation in a unit air column. However, in Eq. (4), this radiation attenuation is rather small because of radiation absorption, while in Eq. (7) the attenuation produced by scattering is generally substantially greater. In addition, the coefficient $\tilde{f}$ whose magnitude is influenced by the pyranometer sensitivity for all directions can be substantially greater than $\psi_2(\pi - \Theta) - 1$ in Eq. (4), which refers to directions rather far from the horizon.

To make the calculation of the coefficient $\tilde{f}$ more convenient, we analyze it into two components which refer, respectively, to the lower and upper hemispheres, i.e., let us assume the following:

$$\tilde{f} = \tilde{f}_1 + \tilde{f}_2. \qquad (10)$$

The equations for calculating $\tilde{f}_1$ and $\tilde{f}_2$ by numerical integration, according to Eq. (9), can be given the following form:

$$\tilde{f}_1 = 2\pi\sum_{0}^{1}\frac{1}{s}\frac{d\bar{i}}{d\tau}\cos\Theta\bar{f}(\Theta)\Delta(\cos\Theta) = 2\pi\sum_{0}^{1}\Phi'(\Theta)\bar{f}(\Theta)\Delta(\cos\Theta), \qquad (11)$$

$$\tilde{f}_2 = 2\pi\sum_{-1}^{0}\frac{1}{s}\frac{d\bar{i}}{d\tau}\cos\Theta\bar{f}(\Theta)\Delta(\cos\Theta) =$$

$$= 2\pi\sum_{-1}^{0}\Phi'(\Theta)\bar{f}(\Theta)\Delta(\cos\Theta), \qquad (11')$$

where the bar over $\underline{i}$ and f(Θ) denote averaging with respect to azimuths.

Because there are no empirical data on the dependence of the brightness of the upper and lower hemispheres on altitude, we are obliged to use the theoretical values of the derivatives $di/d\tau$ in calculating $\tilde{f}_1$ and $\tilde{f}_2$. Here, since the greatest values of $\bar{f}(\Theta)$ with respect to absolute magnitude are noted for all pyranometers at large incidence angles, i.e., for directions close to the horizon, it is necessary to take into consideration the multiple scattering of light in the atmosphere. Consequently, we calculated the functions $\Phi^{\uparrow}(\Theta)$ and $\Phi^{\downarrow}(\Theta)$ according to the theory of the multiple scattering of light in the atmosphere developed by Ye.S. Kuznetsov and by using the tables given in reference [4].

According to the theory of Kuznetsov:

$$i^{\uparrow} = \frac{s^{\circ}}{\pi} m \int_0^{\tau} \varphi_A(t) e^{-(\tau - t)m} dt + \frac{E_0^{\downarrow}}{\pi} A e^{-\tau m}, \qquad (12)$$

$$i^{\downarrow} = \frac{s^{\circ}}{\pi} m \int_{\tau}^{\tau^{\circ}} \varphi_A(t) e^{-(t - \tau)m} dt, \qquad (12')$$

where $\underline{m}$ is the optical mass of the atmosphere in the direction under consideration; A is the albedo of the underlying surface*; $\varphi_A(t)$ is the function tabulated in reference [4] and is the radiation flow scattered per unit solid angle by the vertical atmospheric column with unit cross section and altitude corresponding to the optical density $\Delta t = 1$; $E_0^{\downarrow}$ is the over-all irradiance of the earth's surface, the table of values of which is also given in reference [4].

Differentiating Eqs. (12) and (12') with respect to τ and bearing in mind the fact that $s = s^{\circ} e^{-(\tau^{\circ} - \tau) m_{\odot}}$, and also bearing in mind the fact that with a spherical indicatrix, the intensity of scattered light is not a function of the azimuth, we find the following:

$$\Phi^{\uparrow}(\Theta) = \frac{1}{s} \frac{d\bar{i}^{\uparrow}}{d\tau} \cos\Theta =$$

$$= \frac{e^{(\tau^{\circ} - \tau) m_{\odot}}}{\pi} \left[\varphi_A(\tau) - m \int_0^{\tau} \varphi_A(t) e^{-(\tau - t)m} dt - \frac{E_0^{\downarrow}}{s^{\circ}} A e^{-\tau m} \right]. \qquad (13)$$

$$\Phi^{\downarrow}(\Theta) = \frac{1}{s}\frac{d\bar{T}^{\downarrow}}{d\tau}\cos\Theta =$$

$$= \frac{e^{(\tau^* - \tau)m_{\odot}}}{s}\left[\varphi_A(\tau) - m\int_{\tau}^{\tau^*}\varphi_A(t)\,e^{-(t-\tau)m}\,dt\right]. \qquad (13')$$

Equations (10, 11, 11', 13 and 13') enable us to calculate the values of $\tilde{f}$ for a number of characteristic cases, although this requires a substantial period of time.

B. The Selection of Parameter Values for Calculations

If the pyranometers are calibrated at normal ray incidences, $\psi_1(\alpha) = 1/F_1(\alpha)$ and $\psi_2(\alpha) = 1/F_2(\alpha)$, where $F_1(\alpha)$ and $F_2(\alpha)$ are the correction factors to the conversion factor, corresponding to the incidence angle α. If the pyranometer is calibrated in a horizontal position with solar-ray incidence angles of α_0 and no correction for the dependence of sensitivity on incidence angles is introduced into the conversion factor,

$$\psi_1(\alpha) = \frac{F_1(\alpha_0)}{F_1(\alpha)}$$

and

$$\psi_2(\alpha) = \frac{F_2(\alpha_0)}{F_2(\alpha)}. \qquad (14)$$

The curves of the functions $F(\alpha)$ are given in the documentation of the pyranometers on the basis of measurements taken in the Central Checking Bureau for Meteorological Instruments. These curves vary widely. Nevertheless, we frequently encounter pyranometers in which the sensitivity increases when α increases from 0 to 60-70°, after which it drops rather sharply to α = 95 to 98°. If the observations of absorption are made with these pyranometers at solar zenith distances ranging from 50-70°, for practically all directions, in accordance with Eq. (3'), $f(\Theta)$ will be >0. Since in most cases $di/d\tau \cos\Theta > 0$ (in the lower hemisphere $di/d\tau > 0$ and $\cos\Theta > 0$, while in the upper hemisphere $di/d\tau < 0$ and $\cos\Theta < 0$), according to Eqs. (9) and

(7), $\tilde{f}$ and $\delta_3 q$ will be positive, i.e., the observations will produce an overestimation of the absorption.

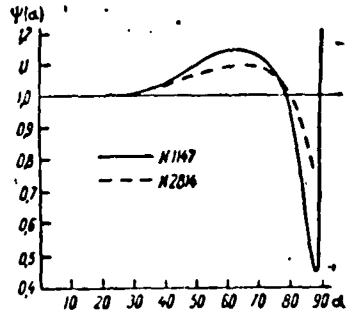

Fig. 1. Dependence of sensitivity of Yanishevskiy pyranometers No. 1147 and No. 2814 on the incidence angle of radiation.

As can be seen from Fig. 1, the indicated characteristics of the dependence of sensitivity on the incidence angle of radiation are very prominent for pyranometer No. 1147, investigated by Yu.D. Yanishevskiy [5] and for pyranometer No. 2814 investigated by Yu.K. Ross [6]. The data for these pyranometers were obtained from numerous calibrations with respect to the sun. In this case, as is pointed out by Ross, the curve for pyranometer No. 2814 given in the certification indicates a much less prominent dependence of sensitivity on α. This forced us to regard the certification curves of F(α) with caution. In addition, these curves are found inadequate, since they break off at $\alpha = 87^{\circ}$, while for our

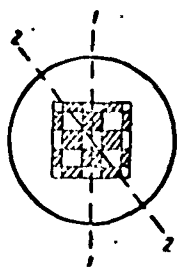

Fig. 2. Yanishevskiy pyronometer with square battery: 1-1 and 2-2 are the directions in which the measurements were made.

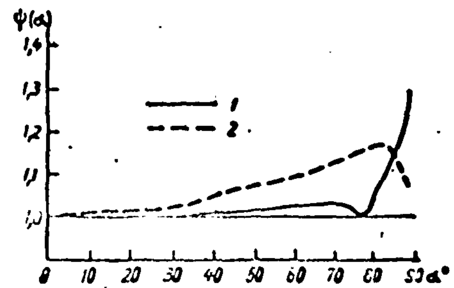

Fig. 3. Dependence of sensitivity on incidence angle of radiation for pyranometer. 1) No. 2661 with battery 4 x 4 cm; 2) No. 2741 with radially symmetrical battery.

calculations we required values reaching at least $\alpha = 115^{\circ}$. Consequently, several different Yanishevskiy pyranometers were investigated for dependence on incidence angles in the Laboratory of Atmospheric Optics and Actinometry of the TsAO*. In this connection, special

attention was given to obtaining sufficiently reliable data for α close to 90° and greater than 90°. One of the important sources of errors in these angles is the illumination of the pyranometer by lamp light which passes over the pyranometer bell, through the pyranometer and is then reflected from the wall of the room or from the screen placed behind the pyranometer. Consequently, in our device, the lamp light passing the pyranometer is reflected by a mirror made of black glass to the reverse side of the diaphragm covered with black velvet. The measurements were made in diverging beams. However, since the distance from the lamp to the pyranometer was greater than 1m, the angles at which the beams fell on the extreme points of the battery differed by no more than 2°. For pyranometers with square batteries, the measurements were made in the directions (azimuths) indicated in Fig. 2 and an average was taken.

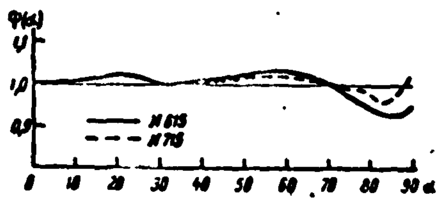

Fig. 4. Sensitivity as a function of incidence angle of radiation for pyranometers with batteries 3 x 3 cm, Nos. 615 and 715.

As is indicated in Figs. 3 and 4, for pyranometers with square batteries, the dependence of sensitivity on incidence angle is very weak. The new pyranometers produced by the Tbilisi Hydrometeorological Instrument factory with 3.5 x x 3.5 batteries were exceptionally good; instruments equipped with them had incidence-angle dependences weaker than indicated by their certificates. The characteristic of radial-symmetrical pyranometer No. 2741 was substantially worse (Fig. 3) than pyranometers with square batteries. An even more undesirable characteristic of these pyranometers is the substantial emf which develops as a result of the illumination of the thermopile by rays reflected from the inner surface of the bell at incidence angles greater than 90°. Since when $\alpha = 90^{\circ}$

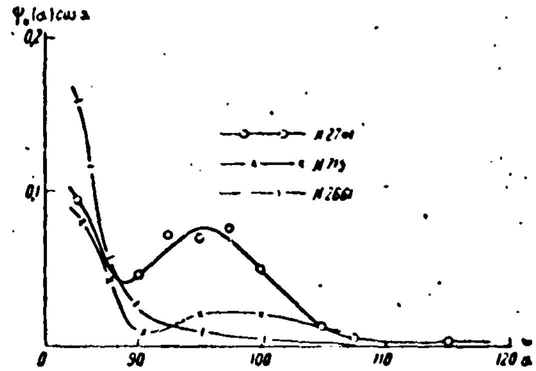

Fig. 5. Dependence of $\psi(\alpha)$ cos α on α for pyranometers Nos. 2741, 715, and 2661 when $85^{\circ} < \alpha < 120^{\circ}$.

$\varphi(\alpha) = \pm \infty$ for angles close to 90°, the pyranometers can be more conveniently characterized not by the magnitude of relative sensitivity $\psi(\alpha) = n(\alpha)/[n(0) \cos \alpha]$, but rather by the magnitude of the relative emf which is $\psi(\alpha) \cos \alpha = n(\alpha)/n(0)$ ($n(\alpha)$, i.e., the pyranometer reading at incidence angle α). The dependence of $\psi(\alpha)$ cos α on α is presented in Fig. 5 for three types of pyranometers for $85^{\circ} < \alpha < 120^{\circ}$. In reference [6], attention has already been given to the most undesirable features of radial-symmetrical pyranometers with respect to sensitivity as a function of the incidence angle.

Since the spectral compositions of the light of an incandescent lamp and the sky differ sharply and the lamp's radiation is primarily concentrated in the infrared region, in order to supplement the measurements described, we also determined the function $\psi(\alpha)$ for one pyranometer with a lamp filtered by SZS-5 glass which cuts the infrared region. Figure 6 shows that within the limits of experimental accuracy, we can, evidently, assume that the function $\psi(\alpha)$ is not a function of wavelength.

Table 1 presents average values of the function $f(\Theta)$ for certain pyranometer pairs and various intervals of the angle Θ. These

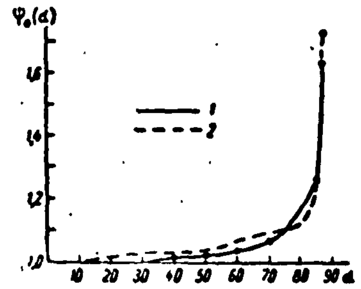

Fig. 6. Pyranometer sensitivity as a function of radiation incidence angle. 1) Without filter; 2) with SZS-5 filter.

values were calculated from Eq. (3') and Figs. 1, 3 and 4. The columns designated as "Upper Pyranometer" and "Lower Pyranometer" indicate the numbers of the pyranometers whose curves of relative sensitivity were used for calculation. For pyranometer No. 1147, for which the relative sensitivity for $\alpha > 90^{\circ}$ is unknown, it was assumed that the sensitivity was the same as in the case of pyranometer No. 2661. It was assumed that the pyranometers with curves of type No. 1147 were calibrated at normal ray incidence. For the remaining pyranometers, it was assumed that they were calibrated in a horizontal position and

TABLE 1

Average Values of Functions f(Θ) in Various Intervals of Angles Θ for Certain Pyranometer Pairs

1 Пиранометр		$h_{\odot}$	Θ								
2 снизу	3 сверху		0–20	20–40	40–50	50–60	60–71	71–82	82–86	86–88	88–90
1147	1147	30	0,13	0,12	0,07	0,01	–0,01	0,12	0,53	0,86	0,65
1147	1147	60	0,01	0,00	–0,05	–0,11	–0,13	0,00	0,41	0,74	0,53
715	2741	30	0,05	0,05	0,04	0,03	0,01	0,12	0,74	1,11	2,50
715	2741	60	–0,02	–0,02	–0,03	–0,04	–0,03	0,05	0,67	1,04	–2,43
2741	715	30	0,05	0,04	0,00	–0,02	–0,05	–0,05	0,06	0,30	–0,78
2741	715	60	0,03	0,02	–0,02	–0,01	–0,07	–0,07	0,04	0,28	–0,80
2661	2661	30	0,03	0,02	0,01	0,01	0,00	0,00	–0,03	0,02	0,20
2661	2661	60	0,01	0,09	–0,01	–0,01	–0,02	–0,02	–0,01	0,00	0,18
1 Пиранометр		$h_{\odot}$	**Θ**								
2 снизу	**3 сверху**		**90–92**	**92–94**	**94–98**	**98–109**	**109–120**	**120–130**	**130–140**	**140–160**	**160–180**
1247	1147	30	0,68	0,90	0,59	0,15	–0,01	0,01	0,07	0,12	0,13
1147	1147	60	0,56	0,78	0,47	0,03	–0,13	–0,11	–0,05	0,00	0,01
715	2741	30	–0,79	0,29	0,05	–0,06	–0,02	0,01	0,03	0,07	0,08
715	2741	60	–0,86	0,22	–0,02	–0,13	–0,09	–0,06	–0,03	0,00	0,01
2741	715	30	2,32	0,91	0,42	0,04	0,01	0,00	0,01	0,02	0,02
2741	715	60	2,31	0,89	0,40	0,02	–0,01	–0,02	–0,01	0,00	0,03
2661	2661	30	0,21	–0,01	–0,03	–0,01	0,00	0,01	0,01	0,02	0,03
2661	2661	60	0,19	–0,05	–0,05	–0,03	–0,02	–0,01	–0,01	0,00	0,01

1) Pyranometer; 2) lower; 3) upper.

that $h_{\odot} = 50^{\circ}$. It should also be noted that for incidence angles

$>90^{\circ}$, small corrections were introduced for the shadowing of the upper hemisphere by the aircraft for the lower pyranometer and of the lower hemisphere for the upper pyranometer. Since these corrections had little effect on the final result of error calculations, we will not devote further time to the method of introducing these corrections.

TABLE 2

Values of Functions of $\Phi^{\uparrow}$ and $\Phi^{\downarrow}$ for Various $\underline{m}$

A	$h_{\odot}$	τ^*	τ	Φ	m: 1	2	4	8	16	32
0,2	30	0.5	0.0	↑	0,118	0,118	0.118	0.118	0,118	0,118
0 2	30	0.5	0.0	↓	0,090	0.030	−0,023	−0,034	−0,024	−0,010
0,2	30	0.5	0.1	↑	0,117	0,107	0,090	0,064	0,033	0,006
0,2	30	0,5	0,1	↓	0.113	0,064	0,010	0,013	−0,016	0.001
0.2	30	0.5	0.2	↑	0.095	0,080	0,059	0 010	0.012	0,000
0,2	30	0.5	0.2	↓	0.102	0,071	0 031	0,000	−0.010	−0,005
0,2	60	0.5	0.0	↑	0.102	0.102	0.102	0.102	0,102	0,102
0.2	60	0.5	0,0	↓	0.104	0.050	0.000	−0,020	−0.016	−0.014
0,2	60	0.5	0.1	↑	0,095	0.078	0,073	0,051	0.025	0,003
0.2	60	0.5	0,1	↓	0,116	0,073	0,026	−0,003	−0.007	0,010
0,8	30	0.2	0.0	↑	0,037	0,037	0 037	0,037	0.037	0,037
0,8	30	0.2	0,0	↓	0.176	0,143	0.095	0,039	−0,033	−0,109
0,8	30	0.2	0,1	↑	0,031	0,029	0.023	0,016	0,007	−0,003
0,8	30	0,2	0,1	↓	0.166	0.151	0,123	0,083	0,036	0,002

The selection of typical curves of $\Psi_1(\alpha)$ and $\Psi_2(\alpha)$ for the lower and upper pyranometers was made on the basis of the following considerations: 1) pyranometer No. 1147 was used because it would give the greatest error because of the dependence of the sensitivity on the incidence angle, as compared to other instruments known to us; 2) pyranometers No. 715 and No. 2741 were used together because a similar pair of pyranometers was used in the summer of 1957 to determine absorption in the Northern Caucasus; 3) pyranometer No. 2661 was one of a series of pyranometers used in 1951-53 in measurement of absorption at Tashkent, Kiev, and Minsk observatories.

Most of the calculations were made for the "summer" variant, which is characterized by high optical atmospheric density ($\tau^* = 0.5$) and low albedo of the earth's surface ($A = 0.2$). The calculations were made for three levels, i.e., for those corresponding to the

values τ = 0.0 (surface of the earth), 0.1, and 0.2. If we orient ourselves to the results of the measurements of the optical density of the atmosphere in the intermediate section of the visible spectrum, i.e., the measurements made in the Laboratory of Atmospheric Optics and Actinometry in the summer of 1957 in the Northern Caucasus, altitudes of approximately 1 and 2 km will correspond to τ = 0.1 and 0.2.

Winter observations of solar-radiation absorption were generally not made. However, in connection with the observations of Shlyakov [11] during the Antarctic summer, it is particularly interesting to consider the errors when the albedo magnitude is great and the atmospheric transmission is high. These errors are also characteristic of our winter. For this "winter" variant, it was assumed that A = 0.8, τ^* = 0.2, τ = 0.0 and 0.1.

We adopted two solar altitudes for the "summer" variant, i.e., 30 and 60°; for the "winter" variant we adopted one, i.e., 30°.

The two functions $\Phi^{\uparrow}(\Theta)$ and $\Phi^{\downarrow}(\Theta)$ calculated for the indicated values of A, τ^*, τ and $h_{\odot}$ according to the tables of Kuznetsov and Ovchinskiy are presented in Table 2. Moreover, the angles Θ are not given as an argument, but rather the "atmospheric masses" m corresponding to these angles. The negative values of $\Phi^{\uparrow}$ encountered in directions very close to the horizon indicate that the brightness of the lower hemisphere in these directions does not increase but rather diminishes with an increase in altitude. By the same token, the negative values of $\Phi^{\downarrow}$ attest to an increase in the brightness of the sky with an increase in altitude.

When we calculate the coefficients $\tilde{f}$ from Eqs. (10, 11 and 11') using the data of Tables 1 and 2, we must know the dependence between the mass m of the atmosphere and the angle Θ. If we confine

ourselves solely to Kuznetsov's theory developed for a flat atmosphere, we must assume that $m = |\sec \Theta|$. However, in so doing we will undoubtedly introduce a rather substantial error for directions close to the horizon. Consequently, in order to make the transition from $\underline{m}$ to angles Θ we use the Bemporad table, although its applicability in the present case has not been strictly proven.* The effective values of coefficients of $\tilde{f}$ presented in Table 3 are to a very great extent a function of the conditions under which the measurements are made. In this case, for the same pair of pyranometers, $\tilde{f}$ can be positive under certain conditions and negative under others. However, the negative values of $\tilde{f}$ for which the result of the measurement of radiation absorption is an underestimation, do not reach values as high as the positive.

The greatest positive values of $\tilde{f}$ and, consequently, the greatest overestimation in radiation-absorption measurement, as one might expect, is given by pyranometers with characteristics similar to No. 1147 and also by a radial-symmetrical pyranometer if it is used in the upper position, when the sun does not stand too high above the horizon.

Table 4 gives an idea of the magnitude of the coefficient $\frac{1}{m_\odot}\frac{ds_1}{dz}$. This table is based on the data in the literature and observations made by the Central Astronomical Observatory during the expeditions of 1951-1952 to Central Asia. The data for the Antarctic was calculated from the results of one flight (6 November 1957) over the Shackleton glacier, which were presented to us by V.I. Shlyakov. The magnitudes included in Table 4 referred to the entire spectrum, including the region of the water-vapor bands. To obtain $\frac{1}{m_\odot}\frac{ds_1}{dz}$, we must introduce into this expression a correction for water-vapor absorption. This correction should be in the order of 00.2 cal $cm^{-2}\cdot min^{-1}/km$. Therefore, when we calculated $\delta_3^1 q$, we adopted the following values:

TABLE 3

λ	$h_\odot$	τ^*	τ	$\tilde{f}$ 1) Пиранометры сверху/снизу				$b_2' q$ 1) Пиранометры сверху снизу			
				1147/1147	2741/715	715/2741	2661/2661	1147/1147	2741/715	715/2741	2661/2661
0.2	30	0,5	0,0	0,117	0,122	−0,035	0,014	0,009	0,010	−0,003	0,001
0.2	30	0,5	0,1	0,076	0,075	−0,005	0,011	0,004	0,004	0,000	0,001
0,2	30	0,5	0,2	0,064	0,051	−0,003	0,008	0,003	0,003	0,000	0,000
0,2	60	0,5	0,0	0,004	0,073	−0,054	−0,003	0,000	0,006	−0,004	0,000
0.2	60	0,5	0,1	−0,036	−0,001	−0,014	−0,008	−0,002	0,000	−0,001	0,000
0,8	30	0,2	0,0	0,060	0,086	—	—	0,001	0,002	—	—
0,8	30	0,2	0,1	0,098	0,030	—	—	0,001	0,000	—	—

1) Upper and lower pyranometers.

TABLE 4

1 Значения, $\frac{1}{m_\odot} - \frac{ds}{dz}$, вычисленные по разным данным

5 Высота, км	2 Средняя Азия		3 Западная Европа		5 Высота, км	4 Антарктида
	6 на горах [7]	7 в свободн. атм.	6 на горах [7]	7 в свобод. атм.		7 в свобод. атм.
0.5—1,5	0,077	0,071	0,103	0,084	0,2—1,4	0,037
1,5—3,0	0,056	0,064	0,064	0,049	1,4—4,3	0,030

1) Values $1/m_\odot - ds/dz$ calculated from various data; 2) Central Asia; 3) Western Europe; 4) Antarctic; 5) altitude in km; 6) in mountains [7]; 7) in free atmosphere.

for the summer variant at $\tau = 0{,}0 \quad \frac{1}{m_{\odot}} \frac{ds_1}{dz} \approx 0{,}08$

. . . . $\tau = 0{,}1$ & $0{,}2 \quad \frac{1}{m_{\odot}} \frac{ds_1}{dz} \approx 0{,}05$

winter $\tau = 0{,}0 \quad \frac{1}{m_{\odot}} \frac{ds_1}{dz} \approx 0{,}02$

. . . . $\tau = 0{,}1 \quad \frac{1}{m_{\odot}} \frac{ds_1}{dz} \approx 0{,}01$

The values obtained were placed on the right side of Table 4. We will discuss them at the end of the article together with errors of other origin. Let us now consider the problem of the effect of scattering asymmetry on $\delta_3^{\prime}q$.

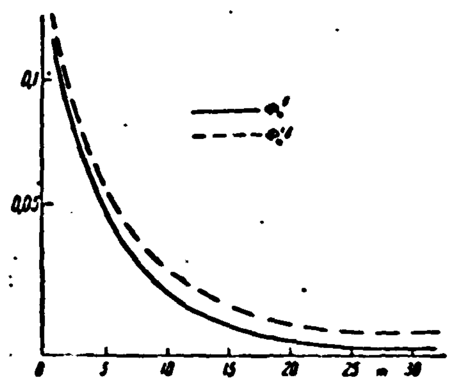

Fig. 7. Functions $\Phi_0^{\downarrow}$ and $\Phi_0^{\prime\downarrow}$ calculated from the tables of Kuznetsov and Ovchinskiy: $h_{\odot} = 60^{\circ}$; $A = 0.2$; $\tau^* = 0.2$.

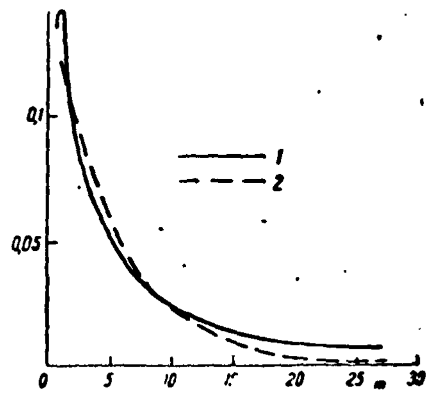

Fig. 8. Functions $\Phi_0^{\prime\downarrow}$ 1) according to data on the brightness of the sky and 2) according to theoretical calculations: $h_{\odot} = 60^{\circ}$; $A = 0.2$; $\tau^* = 0.2$.

The tables of Kuznetsov and Ovchinskiy with which we calculated the functions $\Phi^{\uparrow}$ and $\Phi^{\downarrow}$ correspond to the case of a spherical scattering indicatrix, which is not encountered in the atmosphere. How greatly the calculations diverge from reality as a result of this can be realized, if from function $\Phi^{\downarrow}$ we go to the following function which is close to the above in its physical significance:

$$\Phi_0^{\prime\downarrow}(\theta) = -\frac{1}{S_0}\frac{d\bar{I}_0^{\uparrow}}{d\tau}\cos\theta, \tag{15}$$

where 0 indicates that the corresponding value refers to the earth's surface. Comparing (15) and (13'), and bearing in mind the fact that

in (13') $d\tau = -d(\tau^* - \tau)$ and in both formulas $\cos \Theta < 0$, we can say that both $\Phi^{\downarrow}(\Theta)$ and $\Phi'^{\downarrow}(\Theta)$ characterize the relative changes in the radiation flow proceeding from a certain zone of the sky with the change in optical density of the superjacent atmospheric layer. The difference between these functions is simply that for the function $\Phi^{\downarrow}(\Theta)$, we assume that the optical density changes because of a change in altitude above the earth's surface, while for the function $\Phi'^{\downarrow}(\Theta)$, we assume that the change results from a change in atmospheric turbidity.

Function $\Phi_0'^{\downarrow}(\Theta)$ is easy to calculate from the tables of Kuznetsov and Ovchinskiy, since according to Eqs. (12') and (15),

$$\Phi_0'^{\downarrow}(\Theta) = -\frac{1}{\pi}\varphi_A(\tau^*)e^{-(m-m_\odot)\tau^*}. \tag{16}$$

On the other hand, the empirical values of this function can be obtained from Formula (15) using measurements made on earth of the brightness of the sky and normal illumination by direct sunlight. To do this, we use Reference [8] which gives the values of the visual brightness B at various points of the sky for the following transparency coefficients: $p = 0.87$ ($\tau^* = 0.14$) and $p = 0.80$ ($\tau^* = 0.22$). We assumed a value of 13.5 phots for the solar constant.

As Fig. 7 indicates, the function $\Phi_0'^{\downarrow}$ calculated from Eq. (16) has somewhat higher values for all directions than the functions $\Phi_0^{\downarrow}$. According to Figs. 8 and 9, functions $\Phi'_0{}^{\downarrow}$ obtained from empirical data for very high values of m is even higher. This is a result of the forward distension of the real scattering indicatrix, as a result of which, the attenuation of luminous fluxes coming from the remote layers of the atmosphere is not as great as is the case in symmetrical scattering. The augmentation of $\Phi^{\downarrow}$ for high values of m, as is easy to understand, should lead to an increase in error $\delta_3 q$ for pyranometers

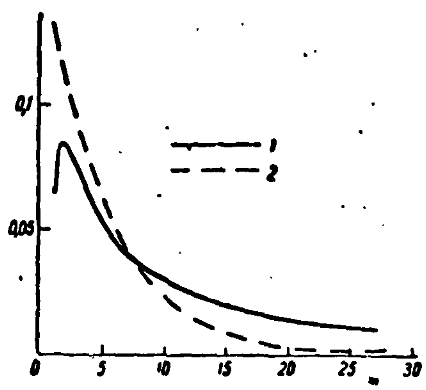

Fig. 9. Function $\Phi'^{\downarrow}$: 1) according to data on the brightness of the sky; 2) according to theoretical calculations: $h_{\odot} = 30^{\circ}$; $A = 0.2$; $\tau^{*} = 0.2$.

with reduced sensitivity at large incidence angles. We make the quite plausible assumption that the changes in the function $\Phi^{\downarrow}$ resulting from scattering asymmetry, are of the same orders as the corresponding changes in function $\Phi'^{\downarrow}$. Then, evidently, the component $\tilde{f}_2$ of coefficient $\tilde{f}$ obtained earlier, which depends on the $\Phi^{\downarrow}$, should be corrected by

$$\delta\tilde{f}_2 = 2\pi \sum_{-1}^{0} \delta[\Phi'^{\downarrow}(\theta)]\,\bar{f}(\theta)\,\Delta(\cos\theta),$$

where $\delta[\Phi'^{\downarrow}(\Theta)]$ is the difference between the values of $\Phi'^{\downarrow}(\Theta)$ for real and symmetrical indicatrices, which can be obtained from Figs. 8 and 9. Thus, we obtain

for $h_{\odot} = 30^{\circ}$ & pyranometers № 1147/№ 1147 $\delta\tilde{f}_2 = -0{,}004$

" " " № 2741/№ 715 $\delta\tilde{f}_2 = -0{,}006$

for $h_{\odot} = 60^{\circ}$ & pyranometers № 1147/№ 1147 $\delta\tilde{f}_2 = 0{,}004$

" " " № 2741/№ 715 $\delta\tilde{f}_2 = 0{,}002$.

For $h_{\odot} = 30^{\circ}$, the correction may even be negative, while for $h_{\odot} = 60^{\circ}$ this correction, although it is positive, is nonetheless so small that it cannot produce substantial changes in the magnitude of the error $\delta_3^{\downarrow}q$.

2. The Effect of a Not Completely Horizontal Pyranometer Set-Up

The effect of change in pyranometer inclination as flight altitude increases was considered earlier [2]. Consequently, we will now assume that the pyranometer inclination is not a function of altitude. In addition, we will assume that the pyranometer sensitivity is not a function of beam direction. The error resulting from pyranometer inclination when we measure radiant-flux density can be broken down

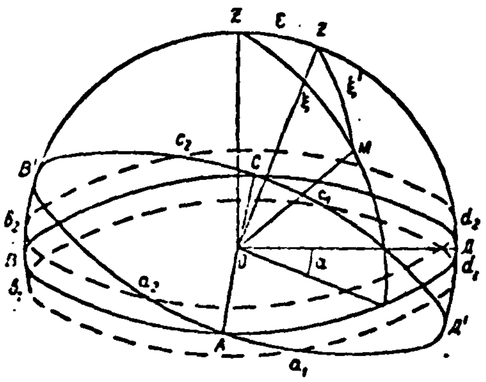

Fig. 10. The effect of pyranometer inclination.

into two components: 1) the error resulting from the fact that the radiation of the sun or some section of the upper or lower hemisphere falls on the receiving surface of the pyranometer not at the angle $\pi - \Theta = \xi$ (zenith distance), but rather at a certain other angle ξ' (Fig. 10); 2) the error resulting from the fact that the radiation of sector ABB'C of the sky near the horizon no longer falls on the receiving surface; in its place, the radiation of sector ADD'C lying beneath the horizontal plane falls on it. Each of these errors is to a great extent a function of the azimuth (relative to the sun) to which the pyranometer is inclined. Let us first of all determine the average errors, assuming that the same number of pyranometer inclinations were encountered in all azimuths in the various flights.

For direct solar-radiation flux the error of the first type in an individual case of inclination at a small angle ε in azimuth a is equal to $S(\cos \xi'_{\odot} - \cos \xi_{\odot})$; moreover, $\cos \xi'_{\odot} = \cos \xi_{\odot} \cos \varepsilon + \sin \xi_{\odot} \sin \varepsilon \cos a$. (For angle symbols see Fig. 10). The average value of this error will be

$$\frac{1}{2\pi}\int_0^{2\pi} S(\cos \xi'_{\odot} - \cos \xi_{\odot})\,da = S\cos \xi_{\odot}(1 - \cos \varepsilon) \approx -\frac{\varepsilon^2}{2} S\cos \xi_{\odot}.$$

Thus, the average value of the relative error of the measured radiant-flux density ($S \cos \xi_{\odot}$), i.e. $-\varepsilon^2/2$, is a constant and does not depend on the solar zenith distance. A similar statement is of course valid with respect to radiation falling on the surface of the upper pyranometer from various parts of the sky and also on the surface of the

lower pyranometer from various sections of the lower hemisphere. Consequently, if both pyranometers are inclined to the horizon at identical angles, the error of the first type for radiation balance will be $\delta_1''(E^{\downarrow} - E^{\uparrow}) = -(\varepsilon^2/2)(E^{\downarrow} - E^{\uparrow})$. Differentiating this equality with respect to altitude and bearing in mind the fact that $(d/dz)(E^{\downarrow} - E^{\uparrow}) = \pm q$, we obtain

$$\delta_1'' q = -\frac{\varepsilon^2}{2} q. \tag{17}$$

Since the average pyranometer inclination in the determination of absorption was always very small and did not exceed a few degrees, this error was negligibly small. Even at an inclination of 6°, the error was only 0.5% q. Under the poorest of circumstances, when only the upper pyranometer was inclined, $\delta_1''q = -(\varepsilon^2/2)(dE^{\downarrow}/dz)$. Since $dE^{\downarrow}/dz$ in the lower section of the troposphere is of the order of 0.1 $cal \cdot cm^{-2}$, min^{-1} km, for $\varepsilon = 6^{\circ} = 0.1$ deg., we obtain $\delta_1''q = 0.0005$ $cal \cdot cm^{-2}$ min^{-1}. Of course this can be entirely neglected.

The elimination of region ABB'C from the number of regions whose radiation is detected by the upper pyranometer and the inclusion of the region ADD'C can be looked upon as a corresponding reduction of pyranometer sensitivity for incidence angles somewhat smaller than 90° and the development or increase of sensitivity for incidence angles somewhat greater than 90°.

Let h denote the absolute magnitude of the angle between the direction of beams and the horizontal plane such that $h = \pi/2 - \Theta$ where $\Theta < \pi/2$ and $h = \Theta - \pi/2$ where $\Theta > \pi/2$.

On the basis of the above we can state that the relative sensitivity of the upper pyranometer for $\pi/2 - \varepsilon_2 < \Theta < \pi/2$ is equal to $\psi_2'(h) = (\cup a_1 d_1 c_1/2\pi)$, while for $\pi/2 < \Theta < \pi/2 + \varepsilon_2$, it is equal to $\psi_2'(h) = 1 - (\cup a_2 b_2 c_2/2\pi)$. Expressing the ares $a_1 d_1 c_1$ and $a_2 b_2 c_2$ in terms of angles ε and h, we obtain respectively for these two cases

$\psi_2^{\downarrow}(h) = -1/\pi \arccos(\mathrm{tg}\, h/\mathrm{tg}\, \varepsilon_2) \simeq -1/\pi \arccos h/\varepsilon_2$ and $\psi_2^{\downarrow}(h) = 1 - 1/\pi \arccos(\mathrm{tg}\, h/\mathrm{tg}\, \varepsilon_2) \simeq 1 - 1/\pi \arccos h/\varepsilon^2$. For the lower pyranometer we can similarly obtain $\psi_1^{\downarrow}(h) = 1 - 1/\pi \arccos h/\varepsilon_1$ for $\pi/2 - \varepsilon_1 < \Theta < < \pi/2$ and $\psi_1^{\downarrow}(h) = -1/\pi \arccos h/\varepsilon_1$ for $\pi/2 < \Theta < \pi/2 + \varepsilon_1$. Consequently, if the pyranometers are calibrated in a horizontal position at approximately the same solar altitude at which the observations are made, in accordance with Eq. (3'),

$$f'(h) = \frac{1}{\pi}\left[\arccos\frac{h}{\varepsilon_1} + \arccos\frac{h}{\varepsilon_2}\right]. \tag{18}$$

Here we must assume that for $h > \varepsilon_1$ $\arccos h/\varepsilon_1 = 0$, and for $h > \varepsilon_2$ $\arccos h/\varepsilon_2 = 0$. Bearing in mind the link between angles Θ and h and substituting (18) in (11) and (11'), we obtain in approximate terms

$$\tilde{f}_1' = 2\left[\int_0^{\varepsilon_1} \Phi^{\uparrow}(h) \arccos\frac{h}{\varepsilon_1}\cos h\, dh + \int_0^{\varepsilon_2} \Phi^{\uparrow}(h) \arccos\frac{h}{\varepsilon_2}\cos h\, dh\right]$$
$$\simeq 2\left[\Phi_{1,cp}^{\uparrow}\int_0^{\varepsilon_1} \arccos\frac{h}{\varepsilon_1}\, dh + \Phi_{2,cp}^{\uparrow}\int_0^{\varepsilon_2} \arccos\frac{h}{\varepsilon_2}\, dh\right] =$$
$$= 2\left(\Phi_{1,cp}^{\uparrow}\varepsilon_1 + \Phi_{2,cp}^{\uparrow}\varepsilon_2\right). \tag{19}$$

Similarly

$$\tilde{f}_2' = 2\,(\Phi_{1,cp}^{\downarrow}\varepsilon_1 + \Phi_{2,cp}^{\uparrow}\varepsilon_2). \tag{20}$$

From Eqs. (18), (19) and (10), we obtain

$$\tilde{f} = 2[(\Phi_{1,cp}^{\uparrow} + \Phi_{1,cp}^{\downarrow})\varepsilon_1 + (\Phi_{2,cp}^{\uparrow} + \Phi_{2,cp}^{\downarrow})\varepsilon_2]. \tag{21}$$

Averaging of functions $\Phi^{\uparrow}$ and $\Phi^{\downarrow}$ when we calculate $\Phi_{1;sr}^{\uparrow}$ and $\Phi_{1;sr}^{\downarrow}$ should be carried out within the limits of angles <u>h</u> from 0 to ε_1, while when we calculate $\Phi_{2;sr}^{\uparrow}$ and $\Phi_{2;sr}^{\downarrow}$, the averaging should be carried out from 0 to ε_2. The error in the absorption intensity resulting from pyranometer inclination can now be calculated from Eq. (7), if we substitute $\tilde{f}$ for $\tilde{f}'$ in this equation.

The inclinations of the upper pyranometer to the horizontal plane during observations were determined repeatedly from the divergence between its readings when the courses were changed. In ordinary

flights according the plan of "Sun to the right" and "Sun to the left," only the inclination in a plane perpendicular to the axis of the aircraft (η_1) is determined. In the Moscow flights of 1949-50 made in PO-2 aircraft, the different values of η_1 were not greater than 4°; in the flights of 1951-52 in LI-2 aircraft, they were not greater than 2°. The following table gives the average values of the angles η_1 (in a plane perpendicular to the axis of the aircraft) and η_2 (in a vertical plane passing through the axis of the aircraft) found from five flights in an LI-2 aircraft made in 1952, according to the plan of "Sun to the right," "Sun behind," "Sun to the left," and "Sun ahead."

altitude, km	1	2	3	5
η_1°	0,8	1,2	0,4	0,4
η_2°	−0,6	0,0	−0,2	0,2

Unfortunately, the setting of the lower pyranometer was always considerably less accurate than the setting of the upper. Nevertheless, its average inclination was probably not greater than 10°.

Functions $\Phi^\downarrow$ for altitudes above the horizon less than 10° ($m > 6$), according to the data of Table 2, are negative in the majority of cases and very small with respect to absolute magnitude. Functions $\Phi^\downarrow$ at the very surface of the earth were of the order of 0.1 (for the "Summer" variant). Assuming that the maximum values of angles ε_1 and ε_2 are respectively 4 and 10° (0.07 and 0.17 rad), we find from Eq. (21) $\tilde{f}'_{maks} = 0.048$ to which corresponds $\delta_2''q = (1/m)(ds_1/dz)\tilde{f}' = 0.05\cdot 0.048 = 0.0024$. As is evident from a comparison with Table 3, the type of error under consideration is substantially less than $\delta_3'q$ for certain pyranometers. Similarly, since $F^\uparrow$ falls rapidly with ascent even at an altitude of 1 km ($\tau \sim 0.1$), $\delta_2''q$ should decrease by several times again.

3. The Effect of Radiation Reflected by the Aircraft

In view of the fact that the pyranometers possess a certain sensitivity to beams falling at an incidence angle greater than 90°, the radiation reflected by the surface of the aircraft should have some influence on the results of the measurements. Figure 11 gives the position of the upper pyranometer on the aircraft in two projections. Radiation reflected by the aircraft and striking a unit area of pyranometer at incidence angles ranging from $\pi - \Theta$ to $\pi - (\Theta + \delta\Theta)$ is evidently equal to $B' \cos \Theta d\omega$, where B' is the brightness of the aircraft surface and $d\omega$ is the solid angle that is subtended by the area cdfe from the center of the receiving surface of the pyranometer. Inasmuch as $d\omega = 2u \sin \Theta d\Theta$, where $u = \angle n Pf = \arccos (b/a \operatorname{ctg} \Theta)$, taking into consideration the pyranometer sensitivity as a function of incidence angle expressed by the function $\psi_2(\Theta)(<0)$, we obtain the error resulting from radiation reflected by the aircraft

$$\delta E^{\downarrow} = -4B' \int_0^{\frac{\pi}{2}} \bar{u}\, \bar{\psi}_2(\Theta) \cos\Theta \sin\Theta \, d\Theta = -2B' \sum_0^1 \bar{u}\, \bar{\psi}_2(\Theta)\, \Delta(\sin^2\Theta).$$

If r is the reflection coefficient of the aircraft surface, then $B' = rE^{\downarrow}/\pi$ and, consequently, finally

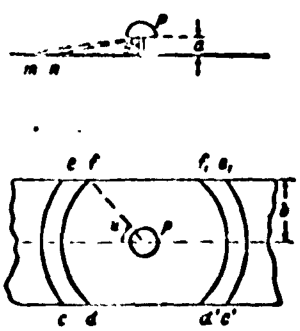

Fig. 11. Effect of radiation reflected from aircraft surface.

$$\delta E^{\downarrow} = -rE^{\downarrow} \sum_0^1 \frac{2\bar{u}}{\pi} \bar{\psi}_2(\Theta)\, \Delta(\sin^2\Theta). \qquad (22)$$

Differentiating Eq. (22) with respect to z and bearing in mind the fact that $(d/dz)\delta E^{\downarrow} = -\delta'' q$, where q, as above, is the radiation-absorption intensity calculated for a layer of unit thickness (1 km), we obtain

$$\delta'' q = -r \frac{dE^{\downarrow}}{dz} \sum_0^1 \frac{2\bar{u}}{\pi} \bar{\psi}_2(\Theta)\, \Delta(\sin^2\Theta). \qquad (23)$$

Table 5 gives the calculation of Σ in

TABLE 5

θ°	60	71	82	86	88	90
$\frac{2\bar{\omega}}{\pi}$	1.00	0.50	0,20	0.10	0.03	
$\bar{\psi}_3(\theta)$	−0.01	−0.10	−0.87	−1.23	−2.68	
$\Delta(\sin^2\theta)$	0.144	0.087	0.014	0.004	0.001	
$\frac{2\bar{\omega}}{\pi}\bar{\psi}_3(\theta)\Delta(\sin^2\theta)$	−0.014	−0.004	−0.002	−0.001	0.000	$\Sigma = -0.023$

the right member of this equality for a radial-symmetrical pyranometer, which, as has already been mentioned, has unfortunate consequences with respect to the dependence of the sensitivity on the incidence angle. Inasmuch as $r \sim 0.5$ and $dE^{\downarrow}/dz$ in the lower layer of the atmosphere is of the order of 0.1 cal·cm^{-2} min^{-1}/km, we obtain $\delta'''q \sim 0.001$ cal·cm^{-2} min^{-1}/km, which is at the very least one order smaller than observed residual absorption.

4. General Conclusions

In accordance with the above, of the errors $\delta_1'q$, $\delta_2'q$, $\delta_3'q$, determined by the dependence of pyranometer sensitivity on the radiation-incidence angle, only $\delta_3'q$ is of practical significance. When pyranometers with a "bad" curve of dependence of sensitivity on incidence angle are used at certain solar altitudes, this value can yield a positive error as high as $\delta_3'q = 0.010$ cal·cm^{-2} min^{-1}/km in the lower part of the troposphere. The corresponding error in the absorption coefficient is about 0.006 km^{-1}. However, such large values of $\delta_3'q$ are observed only in immediate proximity to the Earth. In addition, the lower layer for which the absorption was determined in the majority of cases extends over an altitude ranging from 0.2 to 1.0 km. The average value of τ for this layer is of an order of 0.1 and consequently (in accordance with Table 4) the error $\delta_3'q$ for the lower layer should not be greater than 0.004 cal·cm^{-2} min^{-1}/km.

Of the errors determined for pyranometers which are not entirely

horizontal, only $\delta_2''q$ is of practical significance, but even this error, evidently, even when the instruments have not been set up too carefully, is not greater than 0.002 cal·cm^{-2} min^{-1}/km.

Reflection of solar radiation from the surface of the aircraft also produces a positive error $\delta'''q$ which, however, is of the order of only 0.001 cal·cm^{-2} min^{-1}/km. All of the indicated errors are positive and, consequently, lead to an overestimation of the absorption. We must still add to this the error $\delta_4 q$ due to the effect of bumpy air on the readings of the upper pyranometer; this error is also generally > 0. However, as was indicated in Reference [2], $\delta_4 q$ is a value of a lower order ($\delta_4 q \simeq 0.0004$ cal·cm^{-2} min^{-1}/km). Thus, even in the most unfavorable circumstances, the overall error determined by all of the indicated causes should come to only 0.013 cal·cm^{-2} min^{-1}/km.

Of the errors of other origin, the greatest may be the inaccuracy in determining the temperature coefficient of sensitivity β_t of the pyranometers. Measurement at the Central Astronomical Observatory for a number of pyranometers yielded values varying from −0.04 to −0.10% per degree [2], while at the Bureau for Checking Meteorological Instruments, in the majority of cases somewhat higher values were obtained [9]. Although the average of the values obtained there ($\beta_t =$ $= -0.11$% per degree) is very close to the value (−0.10% per degree) used in processing the observations at a majority of points, it would nevertheless be appropriate to increase this coefficient by a factor of 1.5 in the case of certain pyranometers. For a vertical temperature gradient of approximately 6° per kilometer, this should produce an error in the absorption of the kilometer layer

$$\delta_t q \simeq 0{,}05 \cdot 10^{-2} \cdot 6 \cdot |E^{\downarrow} - E^{\uparrow}| \simeq 0{,}003 \text{ cal·cm}^{-1} \text{ min}^{-1}$$

Closely related to the problem of the effect of change of

pyranometer temperature on the sensitivity of the pyranometer is the problem of the effect of these changes on the magnitude of the parasitic current found when the pyranometers are closed with covers. Since in the case of observations in PO-2 aircraft no covers were used and the zero reading of the galvanometer was read for an open circuit, the parasitic current was not excluded in processing and produced additional errors $\delta_t^! E^{\uparrow}$ and $\delta_t^! E^{\downarrow}$. The magnitude of these errors and the errors corresponding to them in the magnitude of intensity absorbed in the kilometer layer $\delta_t^! q$ could be evaluated using the observations made at Odessa, in which the "zeroes" were read off both with the pyranometers closed and with the circuit open. For the lower layer, $\delta_t^! q$ was on the average -0.0014 cal·cm^{-2} min^{-1}. Although for another pair of pyranometers, this error may be positive, its absolute value is nonetheless sufficiently small [sic] to cause concern. For comparison we cite the fact that the intensity of the "residual" absorption (measured absorption minus calculated water-vapor absorption) on the average for the majority of points at which the measurements were made in the lower kilometer layer is approximately 0.016 cal·cm^{-2} min^{-1}/km. Consequently, the systematic measurement errors that we have considered reach the average value of the "residual" absorption only for the most unfortunate combination. Of course, this least favorable combination is highly improbable. Consequently, it is evident that the "residual" absorption is frequently present in the atmosphere and has a measurable, although small, value. On the other hand, in analyzing the results of the measurements of "residual" absorption, we should bear in mind the large errors of these measurements, for more frequently than not these errors produce an overestimation of the absorption.

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Manuscript Page No.	[Footnotes]
3	For $\alpha > 90^{\circ}$, $\psi_1(\alpha)$ must be assumed negative, since the radiant fluxes falling on the receiving surface of the pyranometer are less than zero, while the pyranometer gives positive readings.
7	In Reference [4], q indicates the albedo of the subjacent surface.
9	All pyranometers which were at one time used to measure absorption should be carefully investigated. This is, however, not possible, since most of them have been damaged or completely ruined.
15	For the upper hemisphere, proof of the applicability of the Emporad function in such a case can be found in Reference [8] on page 35.

Manuscript Page No.	[List of Transliterated Symbols]
22	ср = sr = sredniy = average
23	макс = maks = maksimal'nyy = maximum

CERTAIN FEATURES OF RADIATION PROCESSES
IN THE LOWER TROPOSPHERE OF THE ANTARCTIC

V.I. Shlyakhov

1. Setting Up the Problem and Observational Method

Along with actinometrical observations made on earth, of particular interest should be direct measurement of the elements of radiation balance at various altitudes, in order to completely characterize the radiation features of the troposphere in the Antarctic. Consequently, during the second Continental Antarctic expedition we made actinometrical observations on several sounding-aircraft flights. We measured only the short-wave radiation flows, since there is as yet no approved method for measuring the long-wave radiation fluxes under flight conditions.

To measure short-wave radiation in the atmosphere on an aircraft, we used Yanishevskiy pyranometers and a Yanishevskiy pyrheliometer.

One of the shortcomings of the previously used method for measurement of short-wave radiation on an aircraft is the unreliability of the structure for holding the upper pyranometer in a horizontal position during the flight. In the case of the lower pyranometer, deviation from the horizontal plane of several degrees can be neglected, but for the upper pyranometer this deviation is much more important. To maintain the receiving surface of the pyranometer in a horizontal position, we used a universal joint built into the upper section of the aircraft fuselage (Fig. 1). The central rod of the universal joint

was made with steel and weighs approximately 6 kg. The thin end of the rod comes out through the universal joint. The pyranometer is bolted on to this part of the rod, while a level, threaded screw and heavy nut are fastened to the lower thick section of the rod located inside the fuselage, in order to keep the joint in a horizontal position during the flight of the aircraft.

The lower pyranometer is screwed onto the rod which is tightly fastened to the lower section of the fuselage.

To measure the scattered radiation, the upper pyranometer is covered with a shield during the flight of the aircraft along the "Sun-to-the-right" course. The shield, which measures 13 x 45 cm, rotates about the horizontal axis of the strut, which is fastened to the fuselage skin 30 cm away from the pivot. The shield is controlled by two cables located inside the aircraft (Fig. 1).

Direct solar radiation was measured with the pyrheliometer. The Sun was aimed at manually through an open door of the aircraft during the flight along the "Sun-to-the-right" course. The measurements were made by the total compensation method.

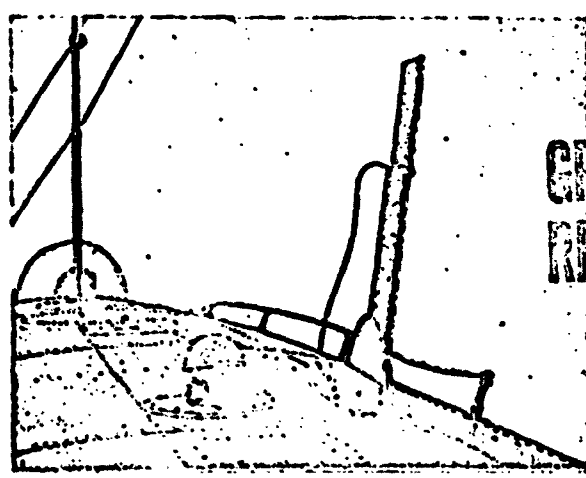

Fig. 1. Set-up of upper pyranometer on the fuselage of LI-2 aircraft.

To measure the emf of the upper and lower pyranometers and also as a zero-galvanometer for measurement with the pyrheliometer, we used a GSA galvanometer, which was mounted on a gimbal suspension. This ensured "zero" stability when the aircraft tilted.

The flights showed that this method of observing the balance components of short-wave radiation and direct solar radiation is worthwhile. Consequently, when we calculated the emf of the upper and lower pyranometers during a very bumpy flight, the galvanometer needle did not oscillate more than one unit. Compensation was also reliably ensured during measurement of direct solar radiation with the pyrheliometer.

When we processed the observations, we introduced corrections into the readings of the upper and lower pyranometers for the dependence of their sensitivity on pressure and temperature. The barometric coefficient of sensitivity was determined in the usual way in a barometric chamber. For both pyranometers used (No. 2203 and No. 2237), the barometric coefficient $\beta_p = -0.5\%$ for 100 mb. The temperature coefficient of sensitivity of the pyranometer was determined from flight observational data by comparing the readings of the sun-illuminated and shaded upper pyranometer with the result of absolute measurements made with the pyrheliometer. Corrections for pressure were first introduced into the pyranometer readings. For both pyranometers, the temperature coefficient was $\beta_t = -0.1\%$ per degree.

Organization of Observations. In the course of 1957 during the 2nd Continental Antarctic expedition for the investigation of the balance components of short-wave radiation in the lower troposphere above various Antarctic regions, 8 flights were made in an LI-2 aircraft. The equipment of the aircraft, in addition to the above-enumerated

instruments, consisted of two meteorographs and aeronavigation instruments.

The observations were conducted on "plateaus" along "Sun-to-the-right" and "Sun-to-the-left" courses at altitudes of approximately 100, 1500, and 4500 m during ascent and descent. To obtain stable radiation characteristics in the free atmosphere, the flights were made in cloudless weather over the even snow-covered surface of the Shackleton ice shelf and the Western glacier in the region of the city Gauss. In addition, regular flights were used to the Pioneer and "Oasis" stations. Glacier reconnaissance flights were also used. Notwithstanding our efforts, it turned out that several flights were made in cloudy weather or at low solar altitudes (less than 20°). Consequently, we succeeded in using only four flights for analysis.

The following measurements were made in each "plateau": 1) altitude according to the altimeter; 2) temperature according to the thermometer on board; 3) galvanometer readings for the upper and lower pyranometers; 4) galvanometer readings for the upper pyranometer when it was shaded, and 5) measurements of direct solar radiation with the pyrheliometer. Scattered radiation was not measured in all the flights. During the flight, a detailed log was kept on the state of the sky and underlying surface. The observations were made by senior scientific coworker V.I. Shlyakhov and engineer-aerologist A.A. Krukovskiy.

2. RESULTS OF OBSERVATIONS

The results of the measurements of the radiation-balance elements and certain other parameters are given in the appendix for each sounding. S (direct solar radiation), $S' = S \sin h$, $E^{\downarrow}$ (density of the descending radiation flux according to the upper pyranometer), $E^{\uparrow}$ (density of the ascending radiation flux according to the lower

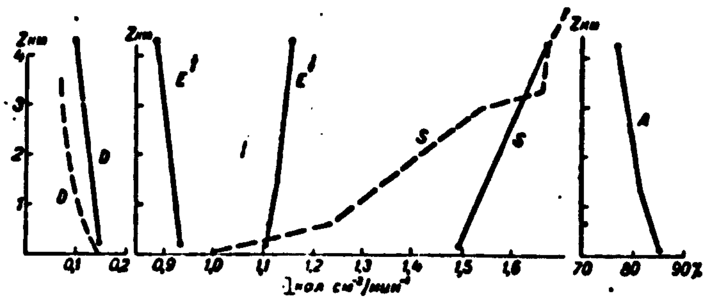

Fig. 2. Balance elements of short-wave radiation as a function of altitude. Shackleton glacier 6/XI/57 h = 39.7°. Dashed lines: D is the density of the scattered-radiation flow according to observations in the Alps; S is the pressure of direct radiation according to observations in an aircraft in Western Europe. 1) cal·cm^{-2}/min^{-1}.

pyranometer), and D (scattered radiation of the sky) are reduced to the same solar altitude, assuming that the increments of S, $E_m{\downarrow}$, $E_m{\uparrow}$, and D_m (m = cosec h) are proportional to the increments of $\sqrt{m}$ [2]. All of the radiant fluxes are expressed in cal·cm^{-2} min^{-1}, using the European actinometrical scale.

When we measured all of the above-indicated radiation fluxes, two values of the short-wave radiation balance were calculated according to the formulas

$$B_1 = E{\downarrow} - E{\uparrow};\ B_2 = S' + D - E{\uparrow}. \qquad (1)$$

With a few exceptions, these values were in satisfactory agreement.

The last column of the appendix gives values of the albedo (A) of the earth's surface in the atmospheric layer lying beneath the level under consideration. The altitude z above the earth's surface is given in kilometers.

Figure 2 gives a general idea of the short-wave balance components as a function of altitude. This is probably satisfactorily typical for a summer day. In this case, for the axis of the abscissas,

the radiation-balance components are given in cal.cm^{-2} min^{-1}, and the albedo is given in per cent. For the ordinate axis, the altitude is given in kilometers. These values are reduced to a mass of 1.56. For purposes of comparison, we have placed the results of observations of other investigators obtained in intermediate latitudes [1b, c, d] on this graph.

As can be seen from Fig. 2 all of the radiation-balance components have a very smooth curve with altitude. From the earth to 4 km, the vertical gradient of direct radiation remains almost constant, i.e., 0.004 cal.cm^{-2} min^{-1} per 100 m, while according to Byutner (dashed line), at altitudes up to 500 m, the vertical gradient of radiation flow is 0.04 cal.cm^{-2} min^{-1} per 100 m and drops to a magnitude of the order of 0.02-0.005 cal.cm^{-2} min^{-1} per 100 m at altitudes greater than 4 km. This difference in the gradient curve of direct solar-radiation flow is explained by the great atmospheric transparency above the Antarctic.

A further examination of Fig. 2 shows that the over-all radiation ($E^{\downarrow}$) increases monotonically, while the reflected radiation ($E^{\uparrow}$) and scattered radiation (D) decrease monotonically with altitude. As far as the albedo is concerned, in contrast to the albedo observed on summer flights in temperate latitudes, it also decreases rapidly with altitude. As Reference [3] indicates, this type of dependence of albedo on altitude should be expected in all cases where the albedo is high.

As can be seen from Fig. 2, the scattered radiation in the Antarctic is greater at all altitudes than in Europe at the same solar altitude. At any rate, this divergence can partially be ascribed to the very great albedo of the underlying surface in the Antarctic.

3. ABSORPTION OF SOLAR RADIATION IN THE ATMOSPHERE

Radiation heating of the air is calculated in degrees per hour from the formula

$$\theta = \frac{245\,\Delta B}{P_1 - P_2}, \qquad (2)$$

where ΔB is the increment in the short-wave radiation bands during transition from the level with pressure P_1 to the level with pressure P_2 [1d]. In accordance with the two methods of calculating the short-wave radiation balance, Table 1 gives two values of θ, denoted as θ_1 and θ_2.

Inasmuch as similar measurements of absorption were made at various points of the USSR for the lower 3 km and gave about 0.1 degree per hour, the values of θ given in Table 1 are a surprise to us since they are seldom so high. Actually, since the presence of smoke and dust in the Antarctic atmosphere is excluded, and the water-vapor content is very small as compared with that found during the summer in intermediate latitudes, it would be natural to assume lower values of θ. However, in the cases under consideration there were also other factors which might increase the radiation heating. In the first place, we must note the very rapid drop in humidity with altitude, as a result of which solar radiation not yet depleted in the water-vapor band region falls into the lower layers of the atmosphere. In addition, the absorption of the intensive flow of reflected radiation should be of greater importance than in the intermediate latitudes.

The heating due to radiation absorption by water vapor was calculated for two roughly systematic distributions of specific humidity (q) with respect to altitude, during the spring and summer (see Table 2). Bearing in mind the unreliability of the data on humidity obtained by an aircraft meteorograph under low-temperature conditions

TABLE 1

Radiation Heating of the Air in the Antarctic During 1957

1 Дата	2 Место зондирования	φ 3 град.	φ 4 мин.	λ град.	λ мин.	z_1	z_2	m	Θ_1	Θ_2
26 X	5 Западный ледник	66	30'	89	00'	0.34	1.38	2.37	0.11	0.06
26/X	" "	66	30	89	00	1.38	4.21	2.37	0.05	0.02
27/X	6 Ст. Пионерская и промежуточный пункт	69 68	40 10	95 94	40 00	0.02	1.48	1.82	—	0.03
6/XI	7 Ледник Шеклтона	66	20	97	00	0.16	1.43	1.74	0.11	0.10
6/XI	" "	66	20	97	00	1.43	4.30	1.74	0.04	0.05
6/XI	" "	66	20	97	00	0.16	1.41	1.56	0.07	0.05
6/XI	" "	66	20	97	00	1.41	4.30	1.56	0.06	0.06
2/XII	" "	66	20	97	00	0.04	1.38	1.44	0.09	—
2/XII	" "	66	20	97	00	1.38	4.21	1.44	0.05	—
2/XII	" "	66	20	97	00	0.04	1.33	1.64	0.05	—
2/XII	" "	66	20	97	00	1.33	4.18	1.64	0.03	—

Note. z_1 and z_2 are the limits of the layer in km above the earth's surface.

1) Date; 2) sounding area; 3) deg; 4) min; 5) Western glacier; 6) Pioneer station and intermediate point; 7) Shackleton glacier.

TABLE 2

Calculation of Air Heating Resulting from Absorption of Solar Radiation by Water Vapor

p	q	w	$(\Delta Б_{H_2O})_{пр}$	$(\Delta Б_{H_2O})_{отр}$	$\Delta Б_{H_2O}$	Θ_{H_2O}
1 Весна $m = 2,38$						
940	0.30	0,050	0.013	0,004	0,017	0,04
820	0,05	0,013	0,025	0.001	0.026	0.02
560		0.000				
2 Лето $m = 1,54$						
970	1,00	0,18	0,029	0,013	0.042	0.07
830	0,15	0,04	0.046	0,003	0,019	0,04
560		0.00				

Note. p is pressure, mb; q is the specific humidity, g/kg; w is the water-vapor content above the level considered, g-cm^{-2}; $(\Delta B_{H_2O})_{pr}$ is the direct radiation flow absorbed by water vapor in the layer; $(\Delta B_{H_2O})_{otr}$ is the reflected radiation flow absorbed by water vapor in the layer; $\Delta B_{H_2O} = (\Delta B_{H_2O})_{pr} + (\Delta B_{H_2O})_{otr}$ (ΔB in cal·cm^{-2} min^{-1}); Θ_{H_2O} is the radiation heating produced by water-vapor absorption, deg. per hr.

TABLE 2 (Note continued)

The albedo in the water-vapor band region is assumed to be 80%.

1) Spring m = 2.38; 2) summer m = 1.54.

and very low relative humidities, the humidity distribution with respect to altitude was to a significant extent taken arbitrarily, assuming a very rapid drop in q above 1.5 km. Consequently, the calculated values of Θ represent the maximum possible increase in temperature resulting from water-vapor absorption.

Comparing the corresponding values of Θ and Θ_{H_2O} in Tables 1 and 2, we arrive at the following conclusions. In the spring (26/X), the calculated heating is close to that measured, if we use the second method of processing the observations. We are, therefore, justified in thinking that the twice-as-high values of Θ obtained using the first processing method are the result of some sort of error in the measurement made by the upper pyranometer. In the summer (6/XI and 2/XII), the calculated heating is of the same order as that observed. However, in both the spring and summer, the calculated heating is nonetheless almost always somewhat lower than that measured. This divergence can probably be explained by such factors not taken into consideration as water-vapor absorption of the scattered radiation of the sky and the absorption by ozone and oxygen. In addition, we must bear in mind the fact that the data on the content of water vapor obtained with an aircraft meteorograph have no pretense to great accuracy.

CONCLUSIONS

1) Due to the great purity and dryness of the atmosphere above the Antarctic, the flow densities of direct and scattered solar radiation change very smoothly with altitude and with small vertical

gradients.

2) The density of the ascending flow (reflected radiation) decreases with altitude, which is explained by the high albedo of the underlying surface.

3) Notwithstanding the purity and dryness of the atmosphere, values of radiation heating of the air in certain cases were observed which were of the same order as those found in intermediate latitudes, which is explained by the peculiarities of the distribution of the specific humidity with respect to altitude and the great density of the ascending radiation flux.*

Manuscript Page No.	[List of Transliterated Symbols]
35	пр = pr = pryamoy = direct
35	отр = otr = otrazhennyy = reflected

Manuscript Page No.	[Footnote]
37	Editor's note. As is apparent from the sufficiently numerous measurements made by V.F. Belov in 1958, low radiation heating of the air must be considered typical for the Antarctic.

APPENDIX

Results of Actinometrical Sounding in the Second Antarctic Expedition

z, км	1 p, мб	t, C	2 q, г/кг	S	S'	E↑	E↓	D	Б₁	Б₂	A, %
				3 кал·см⁻²·мин⁻¹							
4 1. 26/X 1957 г. Западный ледник (φ = 66°31', λ = 29°00' В) h = 24,9°; m = 2,37											
0.34	941	−13,4	0,73	1,370	0,576	0.656	0,555	0,107	0,101	0,127	84.6
1,38	820	−16.4	0,15	1.461	0.616	0.712	0,536	0,094	0,166	0,164	76.7
4,21	558	−29,4	?	1,538	0,647	0.745	0,530	0,067	0,215	0,184	71,2
5 2. 27/X 1957 г. Результаты совместной обработки наблюдений при подъеме на ст. Пионерская (φ = 69°40', λ = 95°40' В) и в промежуточном пункте (φ = 68°10', λ = 94°00' В); h = 33,2°, m = 1,82											
0,02	772	−36,8	0,04	1,554	0,853	—	0,765	0,101	—	0,189	—
1.48	582	−30,6	0,05	1.615	0,888	—	0.769	0,090	—	0.209	—
6 3. 6/XI 1957 г. Ледник Шеклтона (φ = 66°20', λ = 90°00' В); h = 34,9°, m = 1,74											
0.16	976	−7,2	0,90	1,410	0,828	0.958	0,830	0,132	0.126	0,130	86.6
1.43	829	−10,4	0.80	1,524	0,886	1,000	0,809	0.116	0,191	0,187	80.9
4.30	566	−23.4	?	1,624	0,930	1,010	0,779	0,094	0,231	0,245	77.2
7 4. 6/XI 1957 г. Там же h = 39,7°, m = 1,56											
0,16	976	−6,5	1,20	1,494	0,958	1,106	0,940	0,148	0,166	0,166	85.1
1,41	832	−9,9	0,85	1,536	0,991	1,152	0,923	0,130	0,209	0,198	81,5
4,30	566	−23,2	0,05	1,670	1,065	1,160	0,890	0,056	0,270	0,271	76,8
8 5. 2/XII 1957 г. Ледник Шеклтона (φ = 66°10', λ = 97°00' В) h = 45,0°, m = 1,44											
0,04	983	−4,0	1,90	1,489	1,035	1,235	0,982	—	0,253	—	79,5
1,38	824	−12,6	0,45	1,571	1,092	1,266	0,950	—	0,316	—	75,1
4,21	563	−26,3	?	1,698	1,180	1,290	0,925	—	0,365	—	71,3
9 6. 2/XII 1957 г. Там же h = 37,6°; m = 1,64											
0,04	982	0,0	2,08	1,432	0,875	1,075	0,829	—	0,246	—	77,1
1,33	828	−13,3	0,56	1,534	0,937	1,082	0,804	—	0,278	—	74,3
4,18	564	−27,2	0,14	1,651	1,010	1,109	0,795	—	0,314	—	71,8

1) p, mb; 2) q, gkg; 3) cal·cm^{-2} min^{-1}; 4) 1. 26/X 1957. Western glacier (φ = 66°31', λ = 29°00' B) h = 24.9°; m = 2.37; 5) 2. 27/X 1957. Results of simultaneous processing of observations during ascent at Pioneer station (φ = 69°40', λ = 95°40' B) and at an intermediate point (φ = 68°10', λ = 94°00' B); h = 33.2°, m = 1.82; 6) 3. 6/XI 1957. Shackleton glacier (φ = 66°20', λ = 90°00' B); h = 34.9°, m = 1.74; 7) 4. 6/XI 1957. Ibid. h = 39.7°, m = = 1.56; 8) 5. 2/XII 1957. Shackleton glacier (φ = = 66°10', λ = 97°00' B) h = 45.0°, m = 1.44; 9) 6. 2/XII 1957. Ibid. h = 37.6°; m = 1.64.

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ABSORPTION OF SOLAR RADIATION IN THE LOWER TROPOSPHERE

V.G. Kastrov

In the case of a homogeneous underlying surface and cloudless weather, when the radiation field in the atmosphere is a function only of one coordinate, i.e., the altitude (z) above the earth's surface, the intensity of absorption of solar radiation is $\Delta B/\Delta z$, where $B = E^{\downarrow} - E^{\uparrow}$. Here $E^{\downarrow}$ is the density of the descending radiation flux made up of the direct solar-radiation flux and the scattered radiation of the sky, and $E^{\uparrow}$ is the density of the ascending radiation flux composed of radiation reflected from the earth's surface and radiation scattered upward by the underlying layers of the atmosphere. Both fluxes should be taken for a horizontal surface. B is generally called the solar-radiation balance at a given level (z). $E^{\downarrow}$ and $E^{\uparrow}$ were measured at various levels in the USSR in various years with Yanishevskiy pyranometers during flights in free balloons [1] and aircraft [2, 3, 4, 5, 6], and also in automatic stratosphere balloons [7]. This article will discuss certain general results of observations made in aircraft directly by the Central Aerological Observatory in the vicinity of Moscow and during expeditions to Central Asia and to the Northern Caucasus. We will also discuss the results of observations made (on the initiative of the Central Aerological Observatory) in several observatories of the Main Hydrometeorological Office. The observations were made only on clear or almost clear days at a solar altitude not less than 20°. Table 1 gives the basic data characteriz-

ing the observational material obtained.

During the initial processing of the data, which took place at each point, several different temperature and barometric coefficients of pyranometer sensitivity were used without justification. This is not fundamental. Consequently, in preparing this article, we introduced corrections such that the results of the processing correspond at all points to one and the same temperature coefficient (β_t = –0.1% per degree) and one and the same barometric coefficient (β_p = –0.3% per 100 mb) [8]. In addition to calculating $\Delta B/\Delta z$ for various (generally kilometer high) layers, we also calculated: a) the intensity at which solar radiation is absorbed by water vapor $(\Delta B/\Delta z)_{H_2O}$. This is calculated by the well-known Meller formula [sic] based on aerological data on atmospheric humidity at various altitudes. In addition to absorption of direct solar radiation, we also took into consideration the absorption of radiation reflected by the underlying surface.

b) The index of residual (dust?) radiation absorption. This is calculated from the approximate formula

$$a = \frac{\Delta B - (\Delta B)_{H_2O}}{\Delta P/g} \frac{\sin h_{\odot}}{(E' + E')_{cp}}, \qquad (1)$$

where $h_{\odot}$ is the solar altitude.

The below-given values of α will be expressed in % per 100 g/cm^3 of air or, which is practically one and the same for the altitudes under consideration, for 1 km.

Since the changes in solar radiation balance with altitude are small, the values of $\Delta B/\Delta z$ are burdened with large random errors. This applies to an even greater extent, of course, to α. As a result of this, α is often negative, which of course has no meaning.

Table 1 gives the average values of α together with the rms errors of α. The latter characterize the accuracy of the determination of α according to the results of observations as well as the

TABLE 1

General Observations of Absorption of Solar Radiation at Various Points

1 Пункт	2 Годы	3 Число вполне удачных полетов*	4 Высота зондир., км	5 Тип самолета	6 Место преимущественного зондирования	7 Коэффициент остаточного поглощения (10^{-2} на км)	
						$z < 1$ км	1 км $< z \leq 4$ км
8 Москва	1949–51	19	3 и 5	22 ПО-2 25 и 24 ЛИ-2	15 В 20 км к северу от Москвы	0,8±0,2	0.7±0.2
9 Ташкент	1951–52	63	3,5	ПО-2	16 В 50–100 км к СЗ от Ташкента	0,9±0.3	0.8±0,1
10 Минск	1951–52–53 1955	14 10	3.5 4	ПО-2 ЯК-12 23	17 В 15 и 60 км к З от Минска	4,9±1.1 –0,2±0,3	0,8±1.0 0.3±0,1
11 Киев	1951–52–53	30	3	ПО-2	18 В 35–40 км на СЗ от Киева	1,0±0,2	1.1±0,1
12 Одесса	1954 1955	9 7	5 5	ЛИ-2 ЛИ-2	19 В 50 км к С от Одессы	1.4±0,8 0,2±0.9	0.4±0 2 0.1±0,2
13 Кара-Кумы и Кызыл-Кумы	1951 и 52	19	4	ЛИ-2	20 В 50 км к С от Ашхабада и в 100 км к З от Ташкента	1,2±0,3**)	0.3±0.1
14 Северный Кавказ	1957	13	6	ЛИ-2	21 В 50 км к С от Минвод	1.3±0.3***)	0,0±0.1

* Those flights were considered successful in which we succeeded in making observations in all the "plateaus" during ascent as well as descent.
** The data refer to a 0.5-1.0 km layer west of Tashkent in 1951.
*** The data refer to a 0.5-1.5 km layer.

1) Point; 2) years; 3) number of completely successful flights*; 4) altitude of sounding, km; 5) type of aircraft; 6) location of most of sounding; 7) coefficient of residual absorption (10^{-2} per km); 8) Moscow; 9) Tashkent; 10) Minsk; 11) Kiev; 12) Odessa; 13) Kara-Kumy and Kyzyl-Kumy; 14) Northern Caucasus; 15) 20 km north of Moscow; 16) 50-100 km northwest of Tashkent; 17) 15 and 60 km west of Minsk; 18) 35-40 km northwest of Kiev; 19) 50 km north of Odessa; 20) 50 km north of Ashkabad and 100 km west of Tashkent; 21) 50 km north of Minvod; 22) PO-2; 23) YaK-12; 24) LI-2; 25) and.

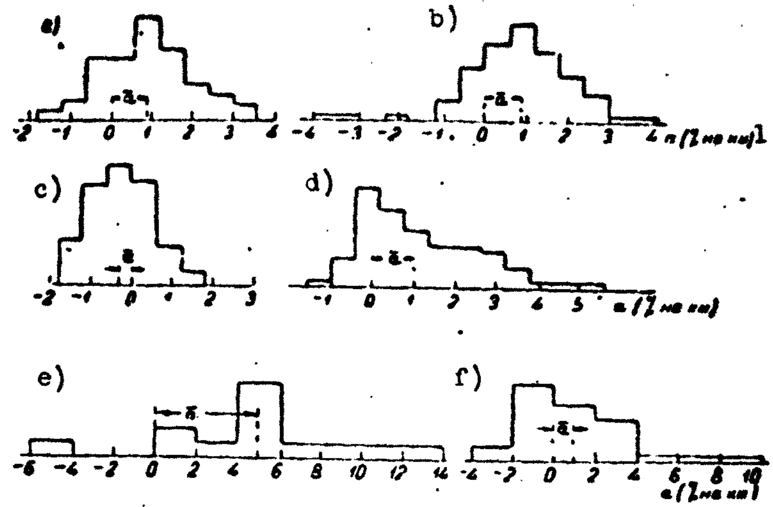

Fig. 1. Recurrence of various values of the index of residual absorption: a) Tashkent, 1952 $z < 3.5$ km; b) Tashkent, 1951 $z < 3.5$ km; c) Minsk, 1955 $z < 4$ km; d) Minsk, 1951-53 $z < 1$ km; e) Minsk, 1951-1953 $1 < z < 3.5$ km; f) Kiev, 1951-53 $z <$ < 3 km. 1) α (% per km).

variability of the residual absorption.

As is apparent from Table 1, in the majority of cases, the average α is close to 1% km^{-1} and changes very little with altitude. Larger deviations from this value are observed only in the following cases: in Minsk in 1951-53 in the lower kilometric layer, where α is 5% km^{-1} (occasional values reached 10-15% km^{-1}), and in Minsk and Odessa in 1955 when the residual absorption was absent for all practical purposes.

Figures 1a, b, c, and d present the recurrence of various values of α at certain points at which observations were conducted for a long period of time.

It should be noted that the curves of recurrence for the observations in the vicinity of Minsk are based on a very small number of cases. They are nevertheless presented by us, since these observations are of particular interest in connection with the enormous values of α encountered here.

According to Fig. 1a and b in Tashkent in 1951 as well as 1952, the recurrence of various values of α was symmetrical with respect to the most frequently encountered value, which was therefore close to the mean. The recurrence of unreal negative values is quite frequent. The symmetrical shape of the curve of recurrence bears witness to the fact that the actual oscillations of the absorption index are small and even in absolute magnitude do not go beyond the limits of the most frequently encountered value.

The curve for the Minsk observations made in 1955 (Fig. 1c) has a similar shape, the only difference being that the average value may even be negative here. Assuming that for all practical purposes there was no residual absroption during the observational period in 1955 in Belorussia, we may assume that the rms error of a single measurement for these observations is $\pm 1\%$ km^{-1}. Undoubtedly the measurement errors were no less at other points and during other periods.

The recurrence curves for observations made in the vicinity of Minsk during 1951-53 (Fig. 1d and e) and in the vicinity of Kiev (Fig. 1f) have an entirely different asymmetrical shape. In this case, for the lower kilometric layer in the vicinity of Minsk, in addition to the maximum with small values of α, a second maximum appears with α of the order of 5% km^{-1}.

As regards the graphs presented, we note further that the maxima of recurrence in the region of negative values of α for certain curves cannot, of course, be explained by random errors alone, but are rather an indication of certain systematic errors in the measurements of the over-all absorption or in the calculations of water-vapor absorption.

In the meteorological literature reference is made to one instance in which an attempt was made to determine the absorption of solar radiation in the atmosphere by a method similar to the one described

above [9]. The author came to the conclusion that the absorption found by him, within the limits of experimental accuracy, agrees with that calculated from the Meller [sic] formula and is produced by the presence of water vapor. Even earlier Waldram [10] attempted to determine the atmospheric absorption of light in the visible section of the spectrum from simultaneous determinations of the atmospheric transmission and scattering indicatrix. In observations on an attached balloon in the vicinity of a large industrial center, he found strong absorption. However, Waldram's results cannot be considered completely reliable, since he did not give sufficient consideration to the aureole effect (Mie effect). The similar but more adequate observations made by Popov [11] in the vicinity of Leningrad gave very close values of attenuation and scattering factors under ground conditions. This is an indication of the absence of absorption in the visible part of the spectrum during observations. As we have seen above, the same was also observed in the free atmosphere in 1955 in Minsk and Odessa. Consequently, the suspicion may arise that the positive values of the coefficient of residual absorption are in fact determined by certain instrumental or methodological errors.

In favor of this we may introduce the following considerations.

1. High residual absorption was observed primarily in the earlier years, when the flights were made in PO-2 aircraft which are much less stable in flight than LI-2 aircraft. The apparatus was in general of lower quality than that used in later years. In addition, it appears that the proper care was not always given to setting up and conducting the observations. In this connection, in Minsk in 1951-53, when such enormous values of α were obtained, the observation materials were much worse than at other points. This is apparent, for example, from the doubtful values of the albedo of the underly-

ing surface, which are obtained from the observational data and which are primarily an indication of the unreliability of the determination of the conversion factor of the lower pyranometer or the fact that the pyranometer has been improperly mounted. In 1955 when the observations were made very carefully, as we have already seen, absolutely no residual absorption was found.

2. The variation in residual absorption on transition from one period to another and from one point to another in many cases cannot be explained by climatic and weather conditions. Consequently, we cannot understand why residual absorption was observed north of Odessa in 1954 and not in 1955. We cannot understand why in the vicinity of Moscow, Tashkent, and Kiev residual absorption was present up to high altitudes, while in the Northern Caucasus and in the vicinity of Odessa (in 1954), residual absorption was found only in the lowest layer of the troposphere.

Very high values of α in the lower kilometric layer in Belorussia can be explained by the effect of the smoke of forest and peat fires sometimes observed here in the warm half year. However, this explanation is not confirmed by the statistics on forest fires, which indicate that for the 1955 absorption observation period we should expect not lower but rather greater values of α than in 1953.

All of these considerations have forced us to consider in greater detail than earlier the various sources of error in the determination of absorption. The results of the analysis are contained in Reference [12]. Although these results indicate that the positive errors in the determination of α should be encountered more frequently than negative errors, it was not possible to entirely attribute to these errors the values of residual absorption obtained in the vicinity of Moscow, Tashkent and Kiev.* What is more, this would be unthinkable

with respect to the very high values of α obtained in the vicinity of Minsk in 1951-53.

Another way of checking the validity of the values of absorption obtained is to investigate the relation to other geophysical values.

Evidently, the average coefficients for the entire sounded layer ($\bar{\alpha}$), if they are valid, should be negatively correlated to the direct solar-radiation intensity for the average distance of the Earth from the Sun for a certain constant atmospheric mass (S_m). It makes sense to try to find this relation by statistical processing of observational data, of course, only when α changes substantially from observation to observation, and there is a large number of observations. In the opposite case, the scattering of radiation and its absorption by water vapor will hide the role of the "residual" absorption. This is evidently what happened in the case of the Kiev observations, which yielded the rms variance $\sigma(\bar{\alpha}) = 0.7\%\ km^{-1}$ and a correlation coefficient between $\bar{\alpha}$ and solar-radiation intensity at $m = 1.5$ of $r = -0.14 \pm 0.26$. For the Minsk observations made in 1951-53 for which $\sigma(\bar{\alpha})$ reached $2\%\ km^{-1}$, the correlation coefficient with radiation intensity at $m = 2$ reaches -0.53 ± 0.21. The rms error is very great here because there are few cases in which both α and S_2 are present (12 cases). We note that a close correlation between α and S would not be expected because the random errors in α are great and the actinometrical observations and observations of absorption were made at various points and, generally speaking, at various times of the day. This latter instance should be of great significance, if the absorbing material is carried by air currents in the form of comparatively small cloud-like masses.

We used the observations made during 1951-53 in Minsk to investigate the relation between the solar-radiation absorption found

and the temperature changes of the lower layers. Initially, according to the data of pyranometrical observations at altitudes of 0.2 and 1.0 km, in addition to the absorption coefficient α, we also calculated the radiation heating Θ expressed in degrees per hour. To do this, we used the formula

$$\Theta = 60 \frac{g}{c_p} \frac{\Delta B}{\Delta p} = 245 \frac{\Delta B}{\Delta p}, \qquad (5)$$

where c_p is the heat capacity of the atmosphere at constant pressure; g is the acceleration of gravity, and Δp is the difference of the pressures at altitudes of 0.2 and 1.0 km. The values of Θ were then compared with the temperature changes at altitudes of 0.22, 0.5, 1.0, 2.0, and 3.0 km above sea level (0.0, 0.3, 0.8, 1.8, and 2.8 km above the Earth's surface) between morning and evening temperature sounding (δT). The latter soundings were carried out at approximately 4 and 16 hours local time. As might have been expected, this comparison yielded a large scattering of different values of δT for similar values of Θ. In addition to the errors in the determination of Θ, this is, of course, also a result of the great influence exerted by a number of other factors on the atmospheric temperature in addition to the influence exerted by radiation heating. Consequently, we used the results of all the observations to clarify the statistical dependence (those observations completed as well as those unfinished but carried out to an altitude of 1 km). Thus, in order to compare with δT, we succeeded in obtaining 25 values of Θ in the layer located at 0.2-1.0 km. Table 2 gives the average values of Θ and δT with their rms errors for three (approximately similar) groups.

According to this table and Fig. 1, δT actually increases continuously in the layer located at 0.2-1.0 km with an increase in the radiation heating in this layer while there is no such dependence be-

TABLE 2

Average Values of Θ in the Layer Located at 0.2-1.0 km and Variation in Atmospheric Temperature Between Morning and Evening Sounding (δT) at Various Altitudes.

Θ	0.07	0.28	0.48
δT [illegible]	11.0±1.0	12.2±0.5	13.4±0.5
δT [illegible]	8.5±1.0	6.7±0.2	7.3±0.7
δT [illegible]	2.3±0.8	3.5±0.3	4.4±0.5
δT [illegible]	1.0±0.8	2.3±0.7	2.1±0.7
δT [illegible]	1.7±0.4	2.0±0.6	1.8±0.6

tween Θ in this layer and δT at high altitudes. Thus, the relation obtained between the radiation-heating rate (Θ) and the actual change in the temperature of the lower layer (δT) of the troposphere confirms the validity of the values of Θ. Although these values are determined not only by the "residual" absorption, but also by water-vapor absorption, the radiation heating resulting from water vapor in the lower kilometric layer is nevertheless not greater than 0.07° per hour and in the majority of cases fluctuates within the narrow limits of 0.04-0.06° per hour. Consequently, the dependence found in Table 2 should almost entirely be attributed to residual absorption. Unfortunately, in this case as in the investigation of the relation between α and S_2, the scattering of the separate values is quite large. This scattering is characterized by the rms errors given in Table 2. For the temperature change at an altitude of 0.8 km above the Earth's surface, we also calculated the correlation coefficient between Θ and δT, which was 0.59 ± 0.15. The correlation coefficient between the residual radiation-heating rate $\Theta\text{-}\Theta_{H_2O}$ and δT is almost exactly equal to the correlation coefficient between Θ and δT.

If when we calculate the correlation coefficients between Θ and δT, we confine ourselves to only 14 cases, when the aircraft-pyranometer observations were made during descent as well as ascent, the

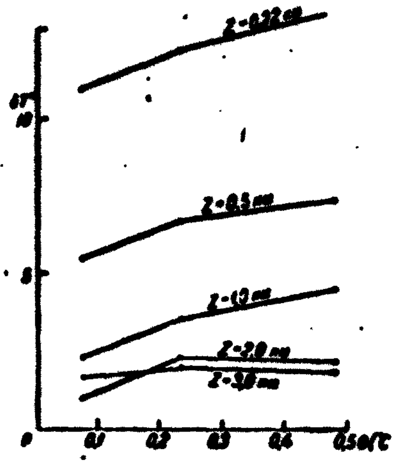

Fig. 2. Temperature change δT between morning and evening sounding as a function of radiation-heating rate Θ in layer located at 0.2-1.0 km. Minsk 1951-53.

correlation coefficient rises to 0.70 ±0.14. The corresponding recording equations have the form

$$\delta T_{1,0} = 2{,}3 + 8{,}6\,\Theta$$
$$\Theta = 0{,}08 + 0{,}57\,\delta T_{1,0}.$$

There has been some surprise about the fact that the "residual" absorption turned out to be even more closely correlated with the temperature changes than with the solar-radiation intensity. This is possibly related to the inadequately representative locale of the actinometric observation station in the city itself.

The conclusion follows from what has been said above that the indices of residual absorption obtained in recent years at various stations, evidently, in general accurately reflect the real radiation processes in the atmosphere, although they are subject to great error. The high values of α obtained in 1951-53 in Belorussia most likely actually were related to forest and peat fires. The recurrence curve of the different values of α, which has two maxima, is in good agreement with this assumption. When $\alpha = 5.10^{-2}$ km^{-1}, the second of these maximums, evidently, corresponds exactly to the appearance of smoky masses of air. The low altitude at which these values of α were distributed indicates that the fires took place comparatively close to the location of the sounding. Because of this, the turbulent mixing cannot transmit the soot particles to a high altitude during the short period of time in which the air is transmitted to the sounding region. The absence of a clear relation

between the recurrence of heavy absorption and forest fires in Belorussia cannot be considered incompatible with the above assumption. This relation might possibly be clarified if we took into consideration the dates of the various fires, their location relative to the sounding area, and the direction of the air flows. In addition, we must also have data on the peat fires often observed in Belorussia, statistical data about which is lacking for the years under consideration.

TABLE 3
Average Values of α 10^2 km^{-1} for Winds at Various Quarters

1 Пункт	2 Направление ветра*			
	С 9	В 10	Ю 11	З 12
3 Москва	0.3±0.4	0.6±0.3	1.1±0.4	1.2±0.5
4 Ташкент	1.0±0.2	0.4±0.2	0.7±0.2	1.0±0.2
5 Минск 1951—53 г.				
s > 1 км	—	0.4±0.4	1.1±1.0	1.0±1.0
s < 1 км	9.0±2.0	6.1±2.3	6.9±1.7	3.8±1.2
1953 г.	—	−0.7±0.2	0.0±0.2	−0.4±0.3
6 Киев	1.0±0.2	1.2±0.3	0.9±0.3	1.4±0.3
7 Одесса	−0.3±0.4	0.3±0.4	0.7±0.6	1.3±0.3
8 Кызыл-Кумы и Кара-Кумы	0.6±0.2	0.3±0.2	0.9±0.4	0.5±0.2

*Direction of wind according to weather vane.

1) Point; 2) direction of wind; 3) Moscow; 4) Tashkent; 5) Minsk; 6) Kiev; 7) Odessa; 8) Kara-Kumy and Kyzyl-Kumy; 9) north; 10) east; 11) south; 12) west.

Above Moscow, Tashkent and Kiev in 1949-53, a considerable amount of residual absorption was quite evenly distributed to an altitude of 4 km. Consequently, it can be assumed that we were concerned here with sufficiently remote sources of absorbing particles. Such sources may also be active in Belorussia, where they add to the effect of the local forces. To clarify the problem of the location of the sources, we present a comparison of the coefficients of residual pressure with wind direction.

Table 3 gives average values of α with their rms errors for different directions of the wind observed in the same layer to which

α refers.

The table indicates that in Tashkent, a somewhat greater residual absorption was observed for the north and west winds blowing from the desert. Evidently, a substantial role is played here by the fine dust raised by the wind from the surface of the desert. In addition to nonabsorbing quartz particles, dark and colored particles are also included in the composition of this dust. Above the desert, during the observational periods (14/IV-10/V 1951 and 15-29/VII 1952), residual absorption was at a minimum for east winds (from the Tashkent Oasis).

The increase in the residual absorption at a number of stations in the European territory of the USSR for west winds is probably to a certain extent related to the contamination of the air by soot and ash particles as the air passes over the countries of Western Europe. We present an elementary calculation which shows that the combustion products given off to the atmosphere in these countries may actually produce increased absorption of solar radiation of an order such as was observed by us for the westerly winds.

As is known, the solid "refuse" from chimneys consists of carbon-soot particles and mineral-ash particles. The former of these have a diameter of 0.25 to 1.0 μ [11]. Since the density of carbon $\gamma = 1.75$ $g.cm^{-3}$, the rate of fall of these particles in unperturbed air according to the Stokes equation (corrected by Millikan) ranges from 0.6 to 9 m per day. Consequently, for all practical purposes these particles can be assumed to be non-settling. If we assume in approximate terms that the effective diameter of a soot particle is 0.4 μ and that the effective solar-radiation wavelength is 0.6 μ, according to the calculations of Ruedy [12], the quantity of scattered and absorbed radiation will be approximately the same; moreover, for one particle per unit volume, the absorption and scattering indices will be α_1 =

$= 1.7 \cdot 10^{-9}$ cm^{-1}. Let us assume that the air travels the course L with the velocity $\underline{v}$ over an industrial section in which on the average Q g/sec of soot is given off into the air for each unit area. If the particles of soot are not drawn off from the atmosphere and are distributed to an altitude of $\underline{z}$, evidently, the air leaving the boundary of the region under consideration will contain on an average QL/vz g/cm^3 of soot in layer $\underline{z}$ or $\frac{QL}{vz\frac{4}{3}\pi r^3} = N$ of particles. For purely approximate calculations we assume that L = 1000 km, v = 10 m/sec, z = = 3 km and r = 0.2 μ. Let us further assume that in the region under consideration, the intensity of soot inflow into the atmosphere is half that found in England. Since in a year $2.42 \cdot 10^6$ tons rise into the atmosphere here and the area of England is $244 \cdot 10^3$ km^2, the value of Q adopted by us is $1.6 \cdot 10^{-11}$ g/cm^2 sec. Substituting the indicated values into the formula for concentration of soot particles, we obtain N = 90 cm^{-3}. Consequently, the radiation absorption index in the calculation for a kilometer will be $\alpha = N\alpha_1 \cdot 10^5 = 1.5 \cdot 10^{-2}$ km^{-1}. Thus, a consideration of only one case of absorption of soot particles provided us with a value of the same order as that observed at a series of stations in the USSR for westerly winds. Our estimation is, of course, probably somewhat exaggerated, since it does not take into consideration the cleansing of particles from the atmosphere during precipitation. In addition, in a number of cases, there may be west winds which have not heretofore passed over a region with heavily developed industry. Finally, we have not taken into consideration the purification of air by smoke-absorbing devices which have recently come into wide use. On the other hand, however, we have not taken into consideration radiation absorption by ash particles. According to Sheleykhovskiy [11], even high-grade coal yields approximately 9% ash, while the mass of soot particles in industrial smoke is only

approximately 1% of the mass of burned out fuel. Consequently, radiation absorption by soot particles can also probably play an important role, although their ability to absorb is of course substantially less than that of soot particles. In addition, some significance should also be attached to the absorption of solar radiation by carbon monoxide, evidently, also of industrial origin, discovered in recent years.

Of course, these calculations and discussions do not entirely eliminate the possibility of the influence of the nonindustrial sources of absorbing particles located to the west of the USSR. As an illustration of the great distances over which the influence of these sources can be spread, we recall that the great forest fires in Canada in 1950 produced substantial atmospheric turbidity in Europe.

Of particular interest and not understood is the complete disappearance of residual absorption in 1955 in Belorussia and in the vicinity of Odessa. Here, although in Belorussia, to a certain extent, the variation in the measured absorption as compared with previous years could be attributed to an improvement in the observational methods and the use of better instruments, in Odessa this factor is completely excluded, since in the period of 1954-55, the observations were made by the same method and even the very same pyranometers were used. We should also discard the assumption to the effect that in 1955 the observations were made primarily for other wind directions. This is evident, for example, from the fact that in the vicinity of Odessa in 1954, the absorption coefficient for northerly and easterly winds was on the average 0.2 and 1.4% per km, while in 1955 for the same winds, it was on the average 0.6 and 0.0% per km. Evidently, further investigations are necessary in order to fully understand the origin of residual absorption.

Let us consider the problem of the possible effect of particles

producing residual absorption on the long-wave radiation balance.

For particles whose dimensions are small as compared with the wavelength the attenuation coefficient is [16]

$$a = \frac{36 \pi w \kappa}{|m^2 + 2|^2 \lambda} m,$$

where $\underline{w}$ is the volume of the particle and $m = n - i\kappa$ is the complex index of refraction. For pure carbon $m = 2 - 2/3\ i$. Therefore, if per 1 cm^2 there are 90 carbon particles, each with a diameter of 0.4 μ, in wavelengths of approximately 10 μ, the attenuation coefficient will be of the order of $0.1 \cdot 10^{-2}\ km^{-1}$. This value is small in comparison with the water-vapor absorption coefficient even in the most transparent region for it at 8–12 μ. The latter, according to Antoni [15] is of the order of 0.05 cm^2/g. For the lower kilometer in the summer, this will correspond to approximately $5 \cdot 10^{-2}\ km^{-1}$ and, consequently, is 50 times greater than the absorption coefficient produced by smoke particles. These factors, evidently, explain the fact that the measurements of the long-wave radiation balance at various altitudes in the free atmosphere do not deviate from the theoretical calculations; this can be attributed to the effect of solid aerosols [17].

In concluding this article, we should note that all of the conclusions arrived at are of a tentative nature. In addition, the causes of certain variations in residual absorption found from the observations are still completely unclear. In the future we consider it desirable in the first place to organize the observations of absorption in the taiga region with parallel detailed recording of all fires.

CONCLUSIONS

1. Observations of solar-radiation absorption at various altitudes above the earth (to 3-5 km) made in recent years at a number of

stations in the USSR indicate that in the majority of cases, the overall absorption is substantially greater than water-vapor absorption derived by calculation from aerological-sounding data.

2. The coefficient of excess absorption α was approximately 1% per km in the majority of cases. Substantial deviations from this value were observed in Belorussia in 1951-53, when in the lower kilometric layer, the average value of α was 5% per km, and, on the other hand, in Belorussia and the vicinity of Odessa in 1955, when residual absorption was completely absent.

3. The existence of strong absorption in Belorussia in 1951-53 is confirmed by the fact that in the majority of cases it is accompanied by a reduction in surface solar-radiation intensity and an augmentation of the increase of atmospheric temperature in the lower kilometric layer of the atmosphere between morning and evening aerological sounding.

4. There is still a great deal that is not clear with respect to the problem of the nature and origin of absorbing particles. Undoubtedly, they are of a different nature at various locations. In Central Asia, they may be the desert dust. In Belorussia in 1951-53, the very high absorption in the lower layer was probably the result of the smoke of forest and peat fires. For many points in the European territory of the Soviet Union, an important role was undoubtedly played by smoke particles transported by winds from Western Europe.

5. Tiny smoke particles which are a source of substantial solar-radiation absorption may not exercise a substantial influence on the balance of the long-wave radiation of the earth and atmosphere.

* * *

In conclusion I should like to express my thanks to the director

of the observatory, Comrade G.I. Golyshev and to assistant director Comrade N.Z. Pinus for their interest in this work and valuable advice which helped in the completion of this work and also to all of the co-workers of the local observatory who took part in the research on absorption.

Manuscript Page No.	[Footnote]
46	Unfortunately, the pyranometers used in previous years were not investigated in detail for dependence of sensitivity on the incidence angle of radiation, which does not permit us to introduce into the data obtained with them corrections based on Reference [12], even though these corrections are of an approximate nature. These corrections would probably be small. This is apparent, for example, from the fact that in Tashkent in 1951 and 1952, the results were on the average in agreement (see Fig. 1a, b), although the observations were made during these years with different pyranometers, while in Odessa in 1954 and 1955, different results were obtained, although the same pyranometers were used.

Manuscript Page No.	[List of Transliterated Symbols]
41	cp = sr = sredniy = average

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THE ASYMMETRY OF LIGHT SCATTERING IN THE ATMOSPHERE

V.G. Kastrov

As is known, the molecular scattering of light and scattering of light by small (as compared with wavelength) particles is symmetrical relative to a plane perpendicular to the beam, i.e., the intensity of the light scattered forward at a scattering angle β is equal to the intensity of the light scattered backward at a scattering angle $\pi-\beta$. On the other hand, the scattering of light by particles whose dimensions are comparable with wavelength is asymmetrical; moreover, more is scattered forward than backward in the majority of cases. Scattering asymmetry is characterized by the asymmetry coefficient,

$$K = I(\beta)/I(\pi-\beta),$$

where I is the intensity of the scattered light. In the case of spherical particles, according to the electromagnetic theory, asymmetry develops with the development of the parameter $\rho = \eta\pi a/\lambda$ where a is the particle radius and λ is the wavelength. Therefore, it would be natural to expect that the scattering asymmetry in any cloudy medium will decrease with an increase in wavelength. However, in determining the light-scattering indicatrix in the atmosphere, a directly proportional relation was found.

Ye.V. Pyaskovskaya-Fesenkova in her monograph on light scattering in the atmosphere [1] touches on this problem, at first indicating that the relation being considered cannot be explained, at any rate completely, by the effect of multiple scattering. After this, she notes

that additional observations are needed in order to understand this phenomenon. This naturally makes the reader think that not even a qualitative explanation of the phenomenon has been found.* But in fact, or so it seems to us, such an explanation is actually contained in the pages following this remark, i.e., 157-161, where the author writes about the effective radius of the particles as a function of the scattering angle. The essence of this explanation consists of the complex nature of light scattering in the atmosphere, resulting from numerous particles of different dimensions. Short-wave scattering is primarily due to small particles (i.e., small as compared with λ) of which there is a greater number than large particles and which scatter short waves much more intensively than long waves. Since these particles scatter light more or less symmetrically, the scattering indicatrices for small λ are not extended forward to any substantial degree. On the other hand, the long waves are very weakly scattered by small particles. They are scattered primarily by large particles, which scatter them as well as the short waves or even better. Indicatrices extended forward markedly are characteristic of large particles; they are, therefore, characteristic for large λ.

In 1946 V.A. Krat [2] derived a formula for the dependence of light-scattering asymmetry on wavelength for an atmosphere consisting of air molecules and large aerosol particles of the same size. Krat's conclusion is in essence only a mathematical expression of the explanation presented by us. His conclusion, however, contains the assumption that the extension of the scattering indicatrix of the aerosol particles themselves are not a function of λ, which is, generally speaking, not true. Therefore, we will present a new derivation of the corresponding formula, which does not contain this assumption.

We will assume that the aerosol particles are spherical. For the intensity of light scattered by one of these particles in some direction, electromagnetic theory yields the expression

$$I = Ea^2 \frac{i(\rho, \beta)}{2\rho^2}, \qquad (1)$$

where E is the illumination of an area perpendicular to the incident rays, and $i(\rho, \beta)$ is a complex function expressed by an infinite series. The values of this function for nontransparent spherical particles are given in Table 16; for drops of water, these values are given in Table 19 and 20 in Reference [3]. For sufficiently small variation intervals of ρ, the function $i(\rho,\beta)/2\rho^2$ can be represented by interpolation formulas of the form

$$\frac{i(\rho, \beta)}{2\rho^2} = F(\beta)\rho^{f(\beta)}, \qquad (2)$$

where in each interval $F(\beta)$ and $f(\beta)$ will have singular values. Table 1 gives values of $f(\beta)$ for scattering angles close to 0 and 180° calculated from the data of the above-indicated tables.

TABLE 1

Values of the Index $f(\beta)$ for Spherical Particles

β \ ρ	1 Непрозрачные шарики			2 Капельки воды				
	0.1–1	1–5	5–10	0.1–1.5	1.5–3	4–6	6–8	8–10
0	4.2	1.7	1.9	5.0	4.0	3.7	1.0	−2.0
10	4.2	1.6	1.0	5.0	3.9	3.1	0.8	−2.4
20	4.2	1.3	−1.6	5.0	3.8	1.6	−0.8	−2.2
30	4.3	0.9	−1.8	5.0	3.2	−1.4	1.3	−0.5
150	3.6	−0.7	0.1	4.1	−0.3	1.2	−0.4	4.2
160	3.6	−0.7	−0.2	4.0	0.2	—	—	4.5
170	3.6	−0.6	−0.3	4.0	−0.1	−4.2	1.2	2.9
180	3.6	−0.6	−0.4	4.0	0.1	—	—	4.9
$\frac{i(180°)}{i(0°)}$ 3 от	8.7	2.1	0.015	1.00	0.075	0.039	—	0.018
4 до	2.1	0.015	0.005	0.065	0.065	—	0.013	0.007

1) Nontransparent spherical particles; 2) drops of water; 3) from; 4) to.

Table 1 indicates that for small particles the backward as well as forward scattering intensity increases rapidly with a decrease in λ (with an increase in ρ). With an increase in the size of the parti-

cles, the dependence of the forward scattering intensity on λ becomes more ~~and more~~ weak, while for large particles ($8 \leq \rho \leq 10$) the nature of this dependence is changed. The dependence of the backward scattering intensity on ρ for $\rho > 1$ is either weak or very irregular. This is, however, not of great significance for the problem under consideration, since for these particles the backward scattering intensity is tens and hundreds of times less than for forward-scattering.

Rayleigh's formula for molecular scattering

$$I = E\frac{c}{\lambda^4}(1+\cos^2\beta) \tag{3}$$

is regarded as a particular case of Eqs. (1) and (2), where, $f_0(\beta) = \text{const} = 4 \text{ и } F_0(\beta) = \frac{c}{(2\pi)^4 a_0^6}(1+\cos^2\beta)$. Here the subscript 0 denotes that the given value refers to molecular scattering. Evidently, $F_0(\beta) = F_0(\pi - \beta)$.

Let us first consider a schematized atmosphere consisting of air molecules and aerosol particles of the same size such that over the entire spectrum under consideration $1 < \rho < 10$ for them. In this case, evidently, the asymmetry coefficient can be expressed by the formula

$$K = \frac{N_0 a_0^3 F_0(\beta)\rho_0^4 + N a^3 F(\beta)\rho^{f(\beta)}}{N_0 a_0^3 F_0(\beta)\rho_0^4 + N a^3 F(\pi-\beta)\rho^{f(\pi-\beta)}},$$

where N means the volume concentration of particles. Bearing in mind the fact that $F(\pi - \beta) \ll F(\beta)$, we reduce this equation to an even simpler form

$$K = 1 + \frac{N a^3 F(\beta)\rho^{f(\beta)}}{N_0 a_0^3 F_0(\beta)\rho_0^4} = 1 + \Phi(\beta)\rho^{-[4-f(\beta)]}, \tag{4}$$

where $\Phi(\beta) = \frac{N a^3 F(\beta)}{N_0 a_0^3 F_0(\beta)}$. Since ρ is inversely proportional to λ, while $f(\beta) < 4$ for the majority of particles, evidently, K should actually increase with an increase in λ. Let us note that this conclusion remains valid if from the expression $1/\rho^2$ in the form of a series of

equations (2) with different values $f(\beta)$ in different ranges of the variation of parameter ρ, we make the transition to the function $f(\beta, \rho)$ which varies continuously with change in ρ. Actually, by substituting $f(\beta)$ for $f(\beta, \rho)$ and denoting $\rho^{-[4-f(\beta,\rho)]}$ in terms of $\underline{x}$, we will have $\frac{1}{x}\frac{dx}{d\rho} = -[4-f(\beta,\rho)] + \frac{\partial f(\beta,\rho)}{\partial \rho}\ln\rho$. Since for large particles $\partial f(\beta, \rho)/\partial\rho < 0$ and $\rho > 1$, $1 \cdot dx/xd\rho < 0$ and, consequently, K decreases on an increase in ρ or with a decrease in λ. It is also interesting to note that despite an observed almost linear dependence of K on λ, according to Eq. (4) the scattering asymmetry should vanish (K should approach 1) only when $\lambda \to 0$.

Ye.V. Pyaskovskaya-Fesenkova showed that the increase in asymmetry with wavelength is characteristic not only for the over-all scattering indicatrix of atmospheric light, but also for the aerosol component of the atmosphere. This can evidently be explained in a way similar to the way in which the dependence of K on λ for the over-all indicatrix was explained. To derive a formula similar to Eq. (4), we will have to assume that there are in the atmosphere two types of aerosol particles to which the small and large values of ρ correspond. Here remembering that over a wide range of variation of $\rho(\rho > 1)$ there is no specific dependence of $f(\beta)$ on λ for all angles close to 180°, it must be assumed that the atmosphere contains a substantial quantity of particles for which $\rho < 1$ or particles for which $\rho > 8$. If we judge from the size distribution curve for aerosol particles proposed by Junge [4], the overwhelming majority of particles actually in the visible region of the spectrum ($\lambda \simeq 0.5\ \mu$) should have $\rho < 1$. There should be few particles with $\beta > 8$. Because of the great asymmetry of light scattering by these particles, they can probably play a role in shaping the brightness of the sky around the sun.

* * *

--- In conclusion I should like to take this opportunity to thank Professor G.V. Rozenberg for the advice he has given me on this article.

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Manu-script Page No.	[Footnote]
62	In the book by K.Ya. Kondrat'yev entitled Radiant Solar Energy, the problem under consideration is considered to be unsolved (see page 100).

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